Changing river styles in response to climate change. Examples from the Maas and Vecht during the Weichselian Pleni- and Lateglacial

Huisink, M.

1999

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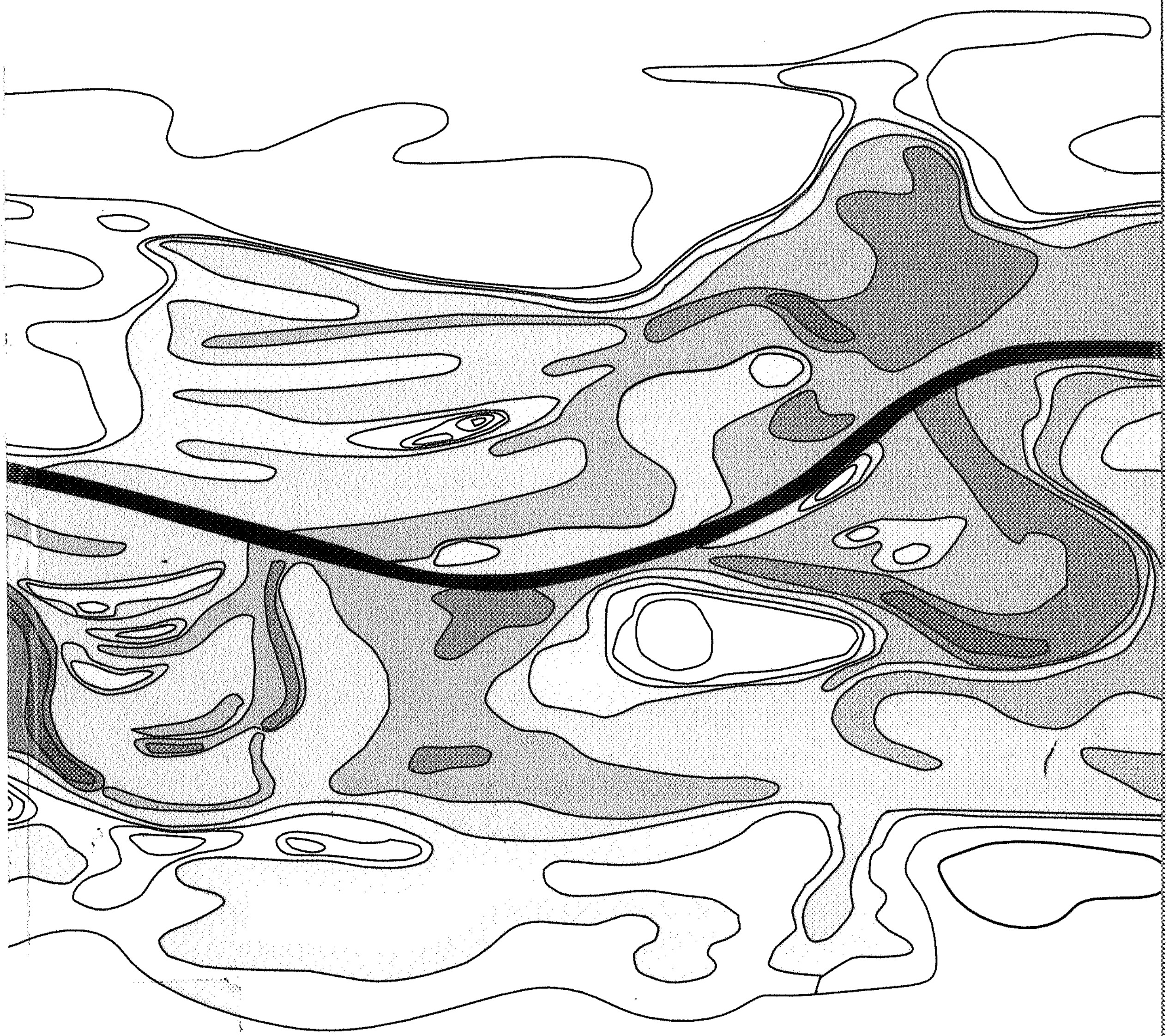

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Margriet Huisink

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CHANGING RIVER STYLES IN RESPONSE TO CLIMATE CHANGE

Examples from the Maas and Vecht during the Weichselian Pleni- and Lateglacial

Cover by Margriet Huisink

This study was carried out at the department of Quaternary Geology and Geomorphology at the Faculty of Earth Sciences, Vrije Universiteit, De Boelelaan 1085, 1081 HV Amsterdam.

The department of Quaternary Geology and Geomorphology participates in the Netherlands Centre for Geo-Ecological Research, ICG.

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CHANGING RIVER STYLES IN RESPONSE TO CLIMATE CHANGE

Examples from the Maas and Vecht during the Weichselian Pleni- and Lateglacial

ACADEMISCH PROEFSCHRIFT

ter verkrijging van de graad van doctor aan
de Vrije Universiteit te Amsterdam,
op gezag van de rector magnificus
prof.dr. T. Sminia,
in het openbaar ter verdedigen
ten overstaan van de promotiecommissie
van de faculteit der aardwetenschappen
op donderdag 11 februari 1999 om 15.45 uur
in het hoofdgebouw van de universiteit,
De Boelelaan 1105

door

Margriet Huisink

geboren te Edam

Promotor: prof.dr. J.Vandenberghe
Copromotor: dr. C. Kasse

Contents 5

Figures

Tables

Tables Appendix

Acknowledgments / Dankwoord

Na afloop van een promotie onderzoek zijn er altijd een hoop mensen die bedankt worden en daar wijk ik niet van af. Hoewel een promotieonderzoek in weze gewoon werk is; een vierjarig contract om wetenschappelijk onderzoek te verrichten, is er bij dit soort werk de afronding met een promotie in een zaal met wijze mensen, voor de gelegenheid zeer apart aangekleed. Er wordt verondersteld dat je als OIO (Onderzoeker In Opleiding) de opleiding tot onderzoeker voltooid hebt. De opleiding bestaat merendeels uit onderzoek doen, waarbij je al doende leert, en uit het volgen van diverse cursussen naar eigen inzicht. In hoeverre dit afstaat van welke student dan ook, die na zijn studie bij een werkgever terecht komt, ingewerkt wordt, en dus al doende leert, op cursussen gestuurd wordt om bij te leren...ik weet het zo net nog niet. Maar goed, in mijn geval heeft mijn werk na 5 jaar geleid tot een promotie. In die 5 jaar ben veel mensen tegengekomen die mij op de een of andere manier geholpen hebben en die ik dan ook wil bedanken.

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1 Introduction

.1 SCOPE

The extensive floods like those in the summer of 1997 in Europe, and the floods in the Maas valley, in 1993 and 1995 show that much has to be learned about fluvial processes and climate. The effect of climate on fluvial systems can be studied by concentrating on the present and historic past (where data sets are available or can be collected) or by focusing on past fluvial responses to past climate changes as an analogue to present or future fluvial behaviour. Both types of studies increase our knowledge on rivers and are therefore very useful. Studies of past fluvial responses provide valuable information on river development on a longer time scale and to climate changes of a larger magnitude. The effect of past climate changes on two Dutch rivers is the subject of this thesis (Fig. 1.1).

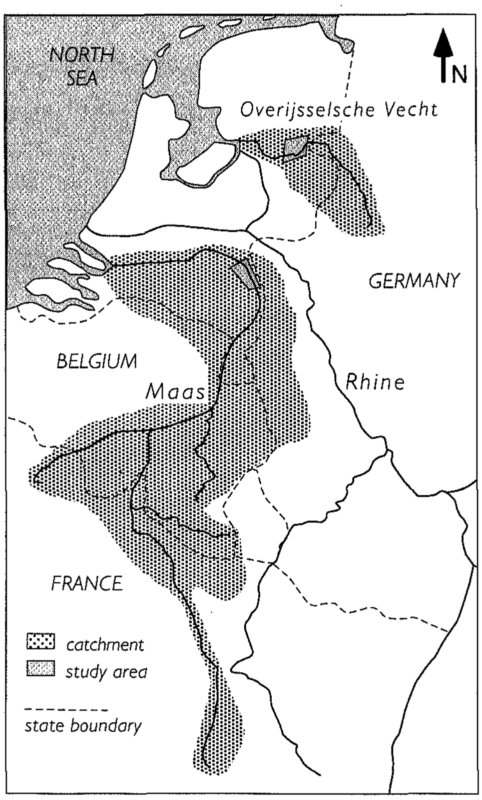


Figure 1.1 Location of Maas and Vecht catchments.

Most attention is paid to the last four periods of well-expressed climate changes, at circa 27 ka, 13 ka, 11 ka and 10 ka (radiocarbon years) which span the late part of the Weichselian Pleniglacial, the whole Lateglacial and the transition to the Holocene (Fig. 1.2 and 1.3). Climate and vegetation reconstructions of these periods in The Netherlands are well documented and dated by, amongst others, Wiegers and Van Geel (1983), Bohncke (1993), Bohncke and Wijmstra (1988), Bohncke et al. (1987), Van Geel et al. (1989), Vandenberghe (1991), Ran and Van Huissteden (1990), Renssen (1997), Hoek (1997a,b), Isarin (1997) and Huijzer and Vandenberghe (1998). This enables a correlation between fluvial and vegetation development which is strongly correlated with climate. The influence of tectonic activity on river style changes in the Maas valley is studied by examining river terrace slopes. Steeper parts in the longitudinal terrace profiles could indicate uplift or subsidence, related to tectonic movements which might influence the channel morphology. Finally, a first attempt is made to quantify amounts of eroded and deposited sediments during the Lateglacial using a GIS. A quantitative approach to fluvial development will increase our understanding of the fluvial system.

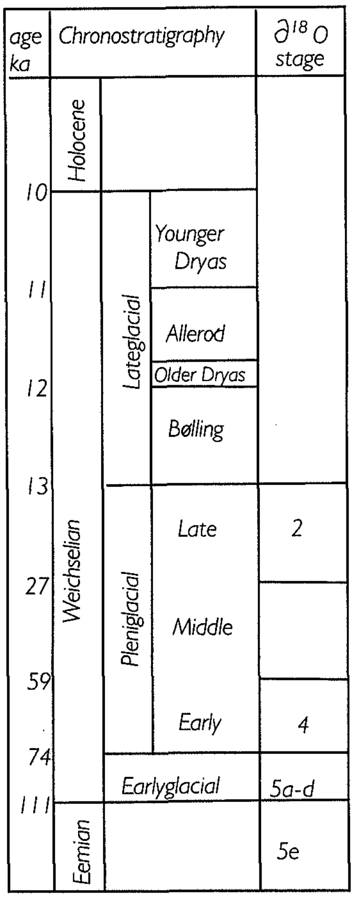


Figure 1.2 Weichselian stratigaphy
(> 30 ka C14 years).

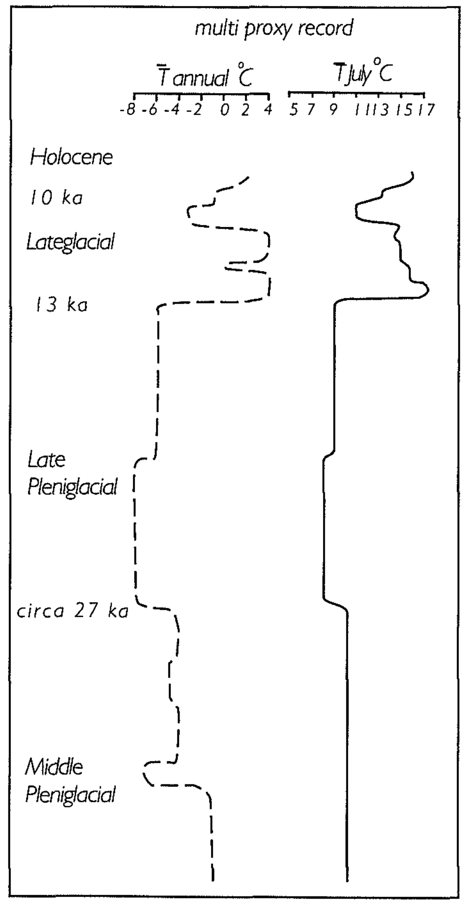


Figure 1.3 Mean July temperature based on palaeobotanical, coleoptera and permafrost evidence. Lateglacial temperature after Hoek (1997a); Pleniglacial temperature after Huijzer and Vandenberghe (1998).

1.2 PREVIOUS WORK

The complex relation between fluvial development and climate-, tectonic- or sealevel changes has been studied in many river basins. The relative influence of each of these factors varies between different river valleys but similarities in fluvial development in Northwest- and Central European lowland river valleys point to the importance of climate change as a triggering factor for fluvial changes. Similarities in fluvial development during the Lateglacial are found for instance in the lower Scheldt valley (Kiden 1991), in lowland Britain (Rose 1995, Collins et al. 1996), in the Ems and Niederrhein valleys (Klostermann 1995) and other German river valleys (Schirmer 1983, Lipps and Caspers 1990), in Polish river valleys like the Warta and Vistula (Vandenberghe et al. 1994, Kozarski 1983, Szumanski 1983, Kalicki and Zernickaya 1995, Starkel 1995) and in the Somme valley in France (Lefevre et al. 1995). Examples of similarities in fluvial development in the Pleniglacial and the transition from Pleni-to Lateglacial are described by Mol

Mol (1997) provides an excellent summary of the changed concepts of the nature of fluvial changes and the causal factors during the last century. Increasingly complex models were postulated to explain fluvial behaviour. Initially the origin of landforms was explained by incision of a river through distinctive stages in response to tectonic uplift (Davis 1899). Later on the effect of climate on rivers became evident by, for instance Büdel (1944), who thought that aggradation occurred during cold periods and incision during warm periods. This is now generally regarded as a too simplistic view. It became evident from Leopold et al. (1964) that a relation exists between bankfull discharge, channel slope and channel pattern. River terraces were formed by two ultimate controls: tectonic forces, that could influence the gradient, and climate forces that affect the hydrologic regime and sediment yield as this is related to vegetation and precipitation. Schumm (1977) pointed out that erosional thresholds should be crossed to invoke a change in the geomorphic system. Extrinsic thresholds relate to external factors, while intrinsic factors are inherent to the system. A special type of intrinsic threshold is the geomorphic threshold which is "the result of landform change through time to a condition of incipient instability, without a change of external influences" (Schumm 1977).

The importance of vegetation cover changes, which relates to climate, on slope erosion rates is large as has been shown by Kirkby (1980). Slope erosion by water diminishes when bare land, shrub, grassland and forests are compared. Forest, and especially needle-leaf trees, is the most erosion free environment due to the high infiltration capacity and percolation (Trimble 1988). Vegetation also strongly influences bank stability (Thorne 1990). A strong forcing of the vegetation cover on river development was suggested by Knox (1984) and Kozarski (1991), as changes in vegetation development and river styles coincided in time. Apart from vegetation there are a number of variables in a fluvial system (time, precipitation, insolation, relief, base level, lithology, geologic structures, human activity, fauna, drainage net, soil profile development etc.) which interact. A change in one or more variables may result in self-enhancing or self-arresting feedback mechanisms (Bull 1991). Internal complex response mechanisms complicate correlations of fluvial responses to climate change within and between physiographic regions (Blum et al. 1994). Rose (1995) emphasized the importance of the location of the valley reach within the catchment, as fluvial responses differ from high relief upstream parts to low relief downstream parts within the catchment.

Increasingly complex models are thus postulated in recent years to explain fluvial responses to climate change. Vandenberghe (1993) proposed a non-linear model in which thresholds were thought to be important in the lead and lag relationships of discharge and sediment supply in response to climate change. Erosion occurred at climatic transitions form cold to warm and vice versa, while aggradation occurred during cold or warm periods. This was explained by delayed vegetation degradation or development (influencing slope and bank stability) and rapid changes in evapotranspiration and runoff. From this theory it is concluded that large changes in fluvial processes occurred during unstable periods when significant climate changes took place.

1.3 STRATEGY

By studying the fluvial development of the Maas and Overijsselsche Vecht changes in geomorphological and sedimentological processes will be determined and used to verify and refine the model of Vandenberghe (1993). By comparing two rivers under similar climate conditions but with different catchments a distinction can be made between local (like tectonic setting) and regional factors that influence fluvial processes. Furthermore, the model is applied for rivers of different sizes. Finally, it is a major objective of the present study to increase our understanding of the nature of thresholds that prevent a river from reacting to climate changes. To compare the fluvial developments of the Maas and Vecht it is necessary to describe the changes in geomorphological and sedimentological processes in detail for each river. Similarities but also differences in fluvial developments between the Maas and Vecht are observed. Perhaps the key question of this study is therefore to wonder how these morphological and sedimentological changes can be explained. Whether they are related to climate change or to other factors like tectonic activity. The tectonic versus climate influence on the develpment of the Maas will be studied in chapter 3 by using palaeo- valley slopes. A correlation between fluvial style - and climate changes will be made in chapters 2 and 6, where the model of Vandenberghe (1993) will also be tested and refined.

Data for this study were collected during two major field works, one for each river valley. During the field work sedimentological data were obtained from mostly hand drillings, arranged in transects across the river valley. Sedimentary structures were studied in small exposures. Since the topography in both river valleys is rather flat, a detailed analysis of 1:10.000 topographical maps was needed to distinguish terrace levels and channels scars. The combination of morphology of channel scars and sedimentology enabled a reconstruction of fluvial style changes which was done in chapter 2 for the River Maas and in chapter 6 for the Overijsselsche Vecht. The terrace flight in the Maas valley shows some striking changes in palaeo-valley slopes. The palaeo-valley slopes were determined by multiple regression analyses. Breaks in these slopes could be explained by tectonic activity in the area. This provides an opportunity to distinguish fluvial responses to tectonic activity from fluvial responses to climate change. In chapter 4 amounts of deposited, eroded or reworked sediment were calculated using a GIS, sediment thickness and terrace surfaces. The amount of reworking gives an indication of the river system dynamics. The net erosion or deposition between successive periods in this part of the Maas valley is calculated as well and it is shown that the architecture of the floodplain probably played an important role in determining the depth of incision. The effect on sediment petrography of the sediment reworking during the Lateglacial by the River Maas and the significant incision in the Younger Dryas is studied in chapter 5 by heavy mineral- and gravel analysis. Finally a compilation of the fluvial development of the Maas and Vecht is given in the synthesis and the observed fluvial responses to climate changes are explained.

1.4 STUDY AREAS

The River Maas, being one of the large rivers in The Netherlands, has been the subject of many geological studies. The first publications about Lateglacial river deposits and river terraces date from the early fifties when Pons (1954) and Pons and Schelling (1951) recognized a "Lower Terrace" level in the Maas valley. A further refinement was made by Pons (1957) who linked fluvial changes of the Maas to climate changes, although he considered sealevel rise as another important factor as well. He distinguished periods of incision and deposition related to warm or cold periods. A division of the Maas valley into five terraces and a recent alluvial plain, based on morphology and soil characteristics, was made by Van den Broek and Maarleveld in 1963. The link with climate changes was not made by them, but their study provided a first stratigraphic framework for the Lateglacial Maas valley. Recently, the fluvial development of the Maas in relation to climatic changes was the topic of papers by Bohncke et al. (1993), Vandenberghe et al. (1994), Berendsen et al. (1995), Kasse (1995b) and Kasse et al. (1995a). The earthquake of Roermond in 1992 generated papers concerning the tectonic activity in the southern part of The Netherlands (Geluk et al. 1994, Van den Berg 1994a, b) and Van den Berg (1996) emphasized the tectonic influence on the Maas evolution.

Few geological studies are known from the Overijsselsche Vecht valley. However, in the early seventies several Dutch geologists and palynologists became interested in a tributary of the Vecht; the Dinkel. The stratigraphy and fluvial evolution of the Dinkel is therefore well known by the work of Van der Hammen (1971 a, b) and Van der Hammen and Wijmstra (1971). Later on the stratigraphy of the Dinkel valley was further refined and modified by Van Huissteden et al. (1986), Van Huissteden and Vandenberghe (1988) and Van Huissteden (1990). In these papers the stratigraphy and changes in sedimentary environment of particularly the Pleniglacial period were described and interpreted. Together with the work of Ran (1990) a thorough fluvial and vegetational reconstruction of the Middle Pleniglacial was made.

The River Maas originates in France, crosses Belgium and enters The Netherlands in the south (Fig. 1.1), while the Vecht starts in Germany and enters The Netherlands in the east (Fig. 1.1). Both rivers are rainfed and discharges may vary considerably through the year. The catchment of the Maas is about ten times as large as that of the Vecht (33.000 km^2 versus 3.700 km^2). The hinterland of the Maas, the Ardennes is subjected to uplift which resulted in a tendency of the Maas to incise. The river crosses several tectonically active horst and graben structures in The Netherlands. The most complete Lateglacial terrace sequence is found where the river flows through the Venlo Graben. The Overijsselsche Vecht valley originated during the Saalian, when ice-masses covered the northern part of the country.Meltwater scoured a wide valley in front of the ice mass which formed the palaeo-Vecht valley. It was filled with sediments after the ice retreat in the Saalian, Eemian and Weichselian.

1.5 FRAMEWORK

This study was financed by the Dutch Organization for Scientific Research (NWO), project 770 07 237, where it was part of the Global Change Program (VvA). The work was carried out at the Faculty of Earth Sciences of the Vrije Universiteit and was incorporated in the 'Netherlands Center for Geoecological Research (ICG)'. This thesis forms part of the research carried out by the department of Quaternary Geology and Geomorphology at the Faculty of Earth Sciences. Within this department the relation between rivers and climate is studied for years and this thesis forms part of the ongoing fluvial research. Previous work by Kasse , Bohncke, Vandenberghe, Van Huissteden and Mol offered a good foundation for the work in this study. Specific hypotheses proposed by them were tested and further elaborated, while in addition the fluvial response to small scale tectonic influences was studied.

2 Late Glacial sedimentological and morphological changes in a lowland river in response to climatic change; the Maas, southern Netherlands

by M. Huisink. Published in Journal of Quaternary Science, vol 12, p 209-223 in 1997

2.1 ABSTRACT

The late Pleniglacial and Late Glacial Maas valley, south of Nijmegen, contains four terraces. Three river systems are described based on the morphology of channel scars on these terrace surfaces and by sediment characteristics. The River Maas reacted to climatic warming at the start of the Weichselian Late Glacial by changing its river system slowly, from a braided system to a transitional phase between braiding and meandering and finally to a highly sinuous meandering system. The Maas reacted rapidly to the Younger Dryas climate deterioration by again establishing a braiding system. At the onset of the Holocene, the river changed abruptly to a meandering river without a transitional phase. The triggering factor for change in the Maas river pattern is almost certainly the changing climate in the Late Glacial. Gradient lines on the terrace surfaces show that tectonic activity did not modify the morphology of the channels. A division of the terraces is shown, the morphological, sedimentological and petrographical characteristics are presented and the linking of changing fluvial patterns with climatic changes or tectonic movements is discussed.

2.2 INTRODUCTION AND GEOLOGICAL SETTING

During the Late Glacial the River Maas repeatedly incised and deposited sediments in a progressively narrowing floodplain. Since the hinterland of the river has been subjected to a long term gradual uplift, it tended to erode and this resulted in the formation of a series of terraces. The repetition of this erosion process over a short period, however, cannot be attributed to steady uplift of the hinterland, neither were base level changes significant since the sea level was some 110 m lower than today (Jelgersma 1966), during the Weichselian.

The Late Glacial terrace stratigraphy is well preserved in the Venlo Graben (Fig. 2.1), in which this study area is situated between Wanssum and Cuijk. The area is the northernmost part of the Maas valley where terrace morphology is still visible. Towards the north-west, Holocene sediments cover the Weichselian morphology: south of Boxmeer incision dominated over deposition and terrace scarps increase in height. Upstream of the Venlo Graben the Maas crosses the relatively higher Peelhorst, where the terrace stratigraphy is only partly preserved.

The Late Pleistocene history of the River Maas is rather well known from previous studies; Pons and Schelling (1951), Pons (1954, 1957, 1966) and Van den Broek and Maarleveld (1963) had previously recognized Late Glacial river systems and terraces. Recently studies by Kasse et al. (1995a), Kasse (1995b), Bohncke et al. (1993), Vandenberghe et al. (1994) and Berendsen et al. (1995) link changes in morphology of river patterns and terrace formation to climatic changes. Another approach has been to explain the forming of terraces by tectonic movements (Van den Berg 1996).

Whether the formation of the Late Glacial terraces is linked to climatic changes and in what manner is the objective of this study. Quantitative data are added to the, mostly qualitative, Maas valley studies. Additional sedimentological data, providing new information on river behaviour, expressed in terms of incision, deposition and stability are supplied and a detailed morphological study is performed. Three Late Glacial terraces, one Pleniglacial and the Holocene floodplain are recognized and correlated with terrace divisions upstream and river system deposits downstream. It will be shown that river pattern changes are not local but occurred over long stretches of the valley at the same time, regardless of locally steeper gradients of the terraces. Heavy mineral analyses are used to describe the terrace sediments and to indicate the decreasing reworking of older material during the Late Glacial. Finally, a model explaining the fluvial reactions on climate changes is validated, using the new information obtained in this study.

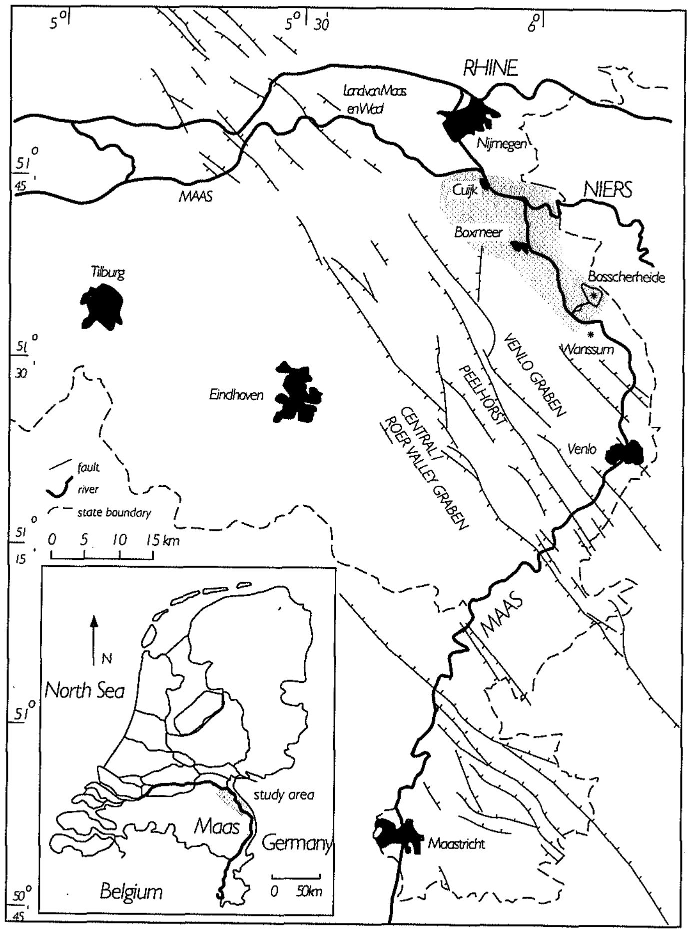


Figure 2.1 Location of study area, faults according to Van Montfrans (1975).

The study area is situated between the Peelhorst in the West, the Saalian ice-pushed ridges in the North and Middle Pleistocene Rhine terraces in the East (Fig. 2.1 and 2.2). South-east of Nijmegen a small tributary, the Niers, discharges into the Maas. This small tributary is unimportant nowadays but on the basis of its relatively large valley dimensions, it was previously of more significance to the Maas. The emphasis in this paper is on the Late Pleniglacial and Late Glacial terraces since these are well developed here, in contrast to the rather narrow Holocene floodplain where more borings would be required to study the Holocene fluvial development in detail which was not the topic of this study.

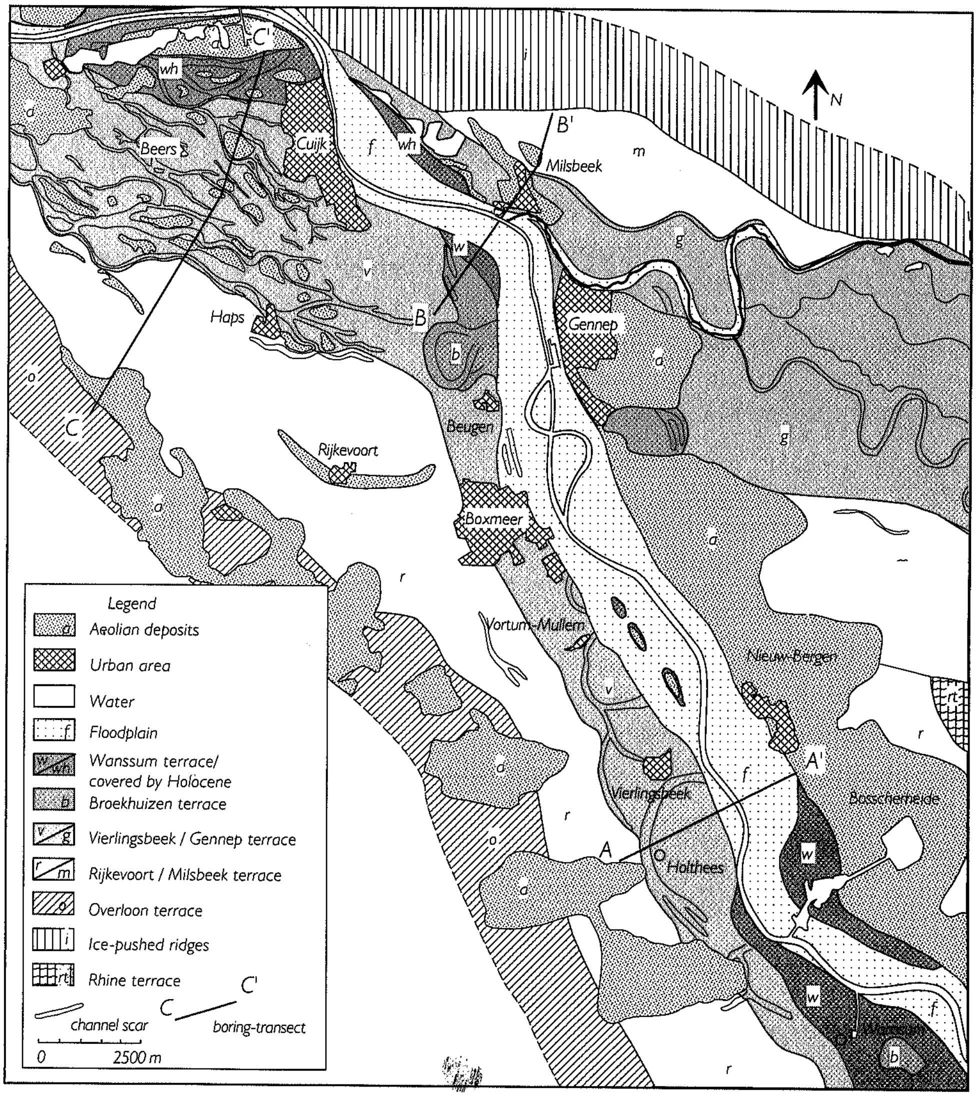


Figure 2.2 Terrace map of Maas valley.

2.3 METHODS

The morphology of the valley is mainly based on detailed analysis of topographical maps (scale 1:10.000). Most of the sedimentological information is obtained from borehole data, individual borings being up to 13 m deep, and arranged in three transects (Fig. 2.2), supplemented by data from exposures. Heavy mineral analysis was performed on the 50 - 420 microns fraction. The grain size classes of sand are based on the Dutch classification (NEN 5104, fine sand: 63 - 150 microns, medium sand: 150 - 300 microns and coarse sand: 300 - 2000 microns). Gradient lines were constructed by using multiple regression analysis.

Chronostratigraphy			*Huisink (this study)*	*Kasse et al. 1995*	*Berendsen et al. 1995*
Holocene		*Subatlantic*			*river system 8 meandering river*
		Subboreal			*river systems 6 + 7 meandering river*
		Atlantic	*Holocene floodplain low-sinuosity meandering river*	*Holocene floodplain low-sinuosity meandering river*	*river system 6 meandering river*
		Boreal			*river system 5 straight/slightly meandering river*
		Preboreal		*level 6 low-sinuosity meandering*	*river system 5 straight/slightly meandering*
Weichselian	*Late Glacial*	*Younger Dryas*	*Wanssum terrace braided system*	*level 5 braided system*	*river system 4 braided system*
		Allerød	*Broekhuizen terrace high-sinuosity meandering river*	*level 4 high-sinuosity meandering river*	*river system 3 high-sinuosity meandering river*
		Bølling	*Vierlingsbeek terrace transitional system*	*level 3 low-sinuosity system*	*river system 2 braided system*
	late Pleniglacial		*Rijkevoort terrace braided system*	*levels 2+1 braided system*	*river system 2 braided system river system 1 braided system*
pre-Weichselian/ Saalian			*Overloon terrace braided system*		

Table 2.1 Comparison of divisions of river systems or terraces of Huisink (this paper), Kasse et al. (1995a) and Berendsen et al. (1995).

2.4 TERRACE STRATIGRAPHY AND RIVER SYSTEMS

The Maas valley contains five terraces, ranging in age from the Late Pleniglacial to the Holocene (Table 2.1, Fig. 2.2), and one older terrace. Each terrace is named after a local village, situated on the terrace surface. This is done to avoid confusion with terrace divisions made by other authors in numbers. Table 2.1 summarizes three divisions in terraces or river systems. The age estimations of the terraces will be explained in the following paragraph.

2.4.1 The Overloon terrace or Peelhorst

This terrace level is part of the Peelhorst (Fig. 2.1 and 2.2), which is a tectonically uplifted area. A clear scarp exists between this level and lower lying younger terraces of the Maas, with a local difference in height of up to seven metres. Two borings (Fig. 2.5) show coarse grained, poorly sorted and gravelly sands with an aeolian sand cover. The sediments were most probably deposited by a braided river system considering the poor sorting and coarse nature of the material.

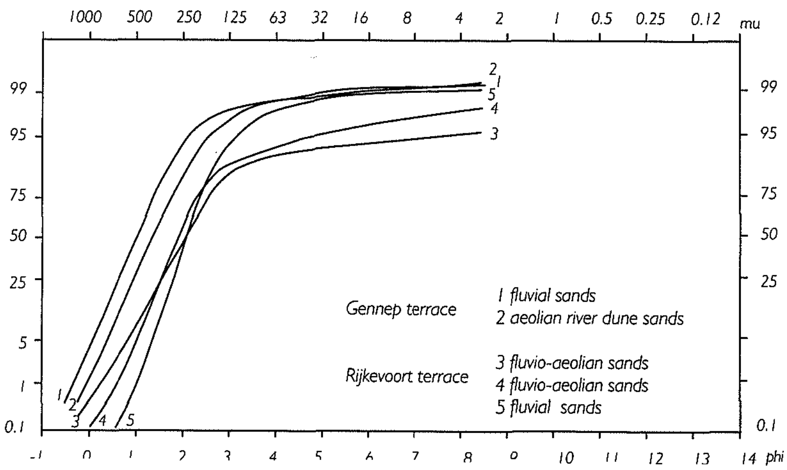


Figure 2.3 Grain-size cumulative frequency curves.

2.4.2 The braided river system of the Rijkevoort and Milsbeek terraces

The surface of the Rijkevoort terrace is rather flat and lacks visible channel scars. The fluvial sediments (Rf1 and Rf2, Fig. 2.5-2.6) are mostly covered by aeolian deposits (Re, Fig. 2.5) with variable thickness (0-4 m). A thick accumulation of aeolian sediments occurs particularly towards the terrace scarp with the higher Overloon terrace. In a small pit at Rijkevoort a gradual transition has been observed from fluvial to fluvio-aeolian deposits immediately beneath the surface. Grain size distributions from these deposits show that both facies have the same mean grain size (210-300 microns, Fig. 2.3). The fluvio-aeolian sands are well sorted in contrast to the fluvial sands that contain clay or silt layers. The sedimentary structures change from planar and trough cross-bedding, locally containing clay pebbles, in the fluvial sands to parallel horizontal and inclined bedding in the fluvio-aeolian sands.

The sediments of the Rijkevoort terrace, on the west side of the Maas, are correlated with the fluvial sediments in the gravel pit Bosscherheide on the east side because of the similar altitude and heavy mineral composition. Two fluvial facies are distinguished at Bosscherheide (Fig. 2.4). The lowermost facies is a poorly sorted, medium to coarse gravelly sand with grain size and bedding types changing rapidly in a vertical and horizontal direction. Channels of 2-5 m wide intersect frequently at short distances and large-scale trough and planar cross-bedding are dominant. The upper facies is a better sorted and less gravelly sand and sedimentary structures have smaller dimensions, with small-scale ripple cross-bedding in 0.5-2 m wide intersecting gullies being dominant. The sediments were deposited by a braided river that was more energetic during deposition of the lower facies. In contrast to the gradual change from fluvial into fluvio-aeolian sands on the west bank, the braided sediments on the east bank are topped by a silt layer (Fig. 2.4) which is an overbank deposit that accumulated during floods in the Bølling, Allerød and Younger Dryas (Bohncke et al. 1993). Involution took place after the coldest phase of the Younger Dryas, 10.800 - 10.500 BP (Kasse 1995 a). During the latest part of the Younger Dryas aeolian sands accumulated in parabolic dunes (upper unit of Fig. 2.4), and it seems that the river shifted its course to the east at this time. The western part of the floodplain became inactive and fluvio-aeolian sands accumulated on top of the fluvial sediments while the eastern part of the river plain was still active.

The morphology and sedimentology of the Milsbeek terrace (Niers river) resembles the Rijkevoort terrace, with gravelly, coarse sands (Fig. 2.6, Mf1) changing upwards to less gravelly medium to coarse sands (Fig. 2.6, Mf2). The absence of aeolian sands and the presence of a rather thick clay layer (Mf3 Fig. 2.6) make it more comparable with the Rijkevoort terrace as it is found in Bosscherheide. The provenance of the sediments from the Milsbeek terrace is different from the Rijkevoort sediments.

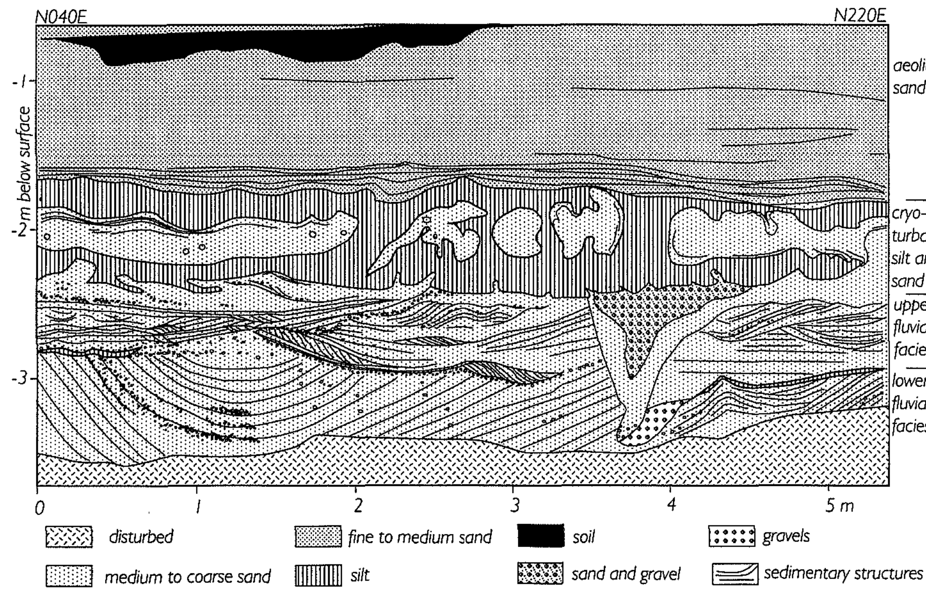


Figure 2.4 Exposure in Late Pleniglacial fluvial sands at gravel pit Bosscherheide.

2.4.3 The transitional system of the Vierlingsbeek and Gennep terraces

Channel scars on the surface of these terraces show characteristics of a transition from a braided to a meandering system. This morphology is distinct, especially near Holthees (Vandenberghe et al. 1994), where several small gullies merge into one large, curved channel. Smaller gullies of the braided system were abandoned while others began to incise up to 3 m and formed small levees (Vf4 Fig. 2.5), so the

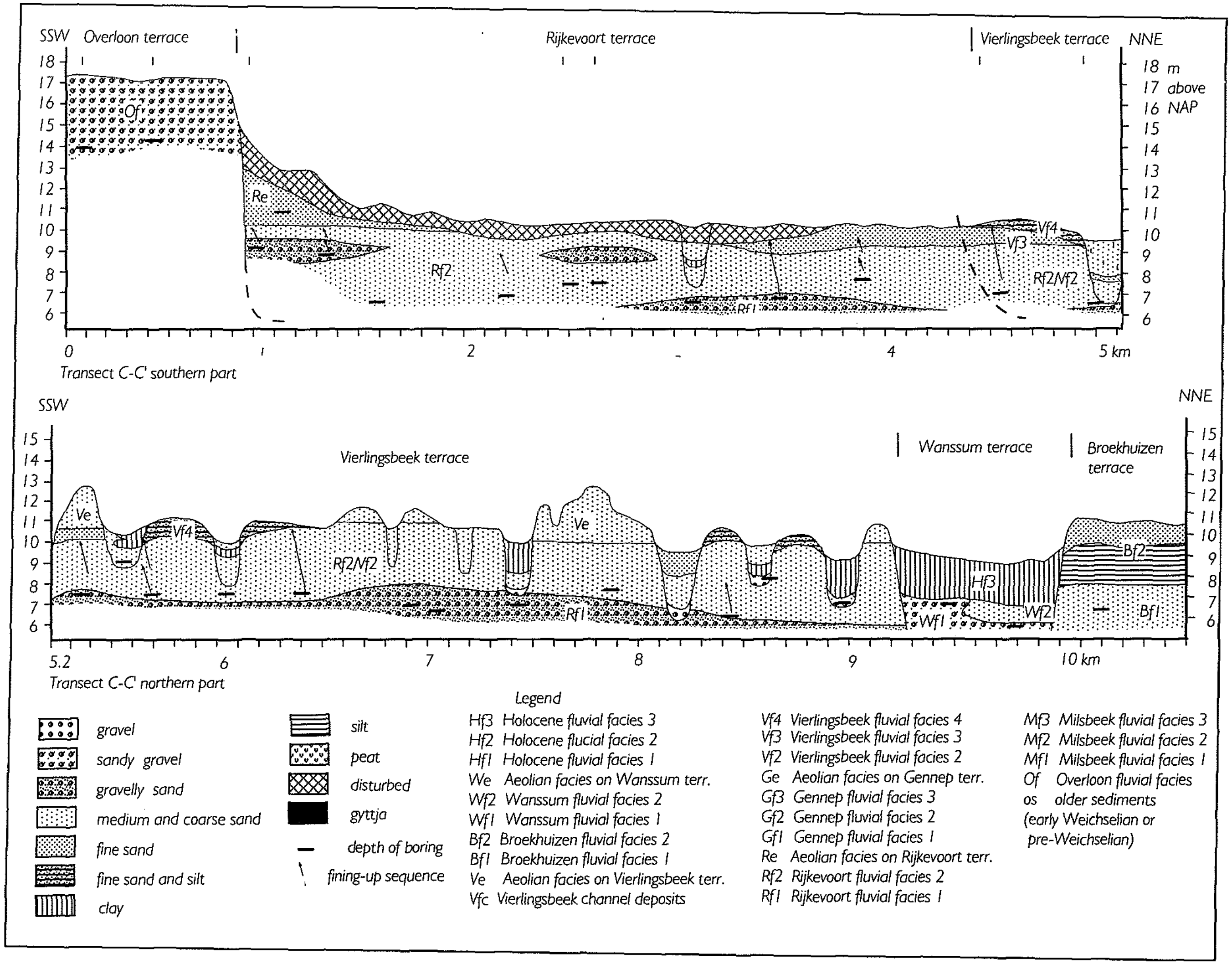


Figure 2.5 Boring transect C-C'.

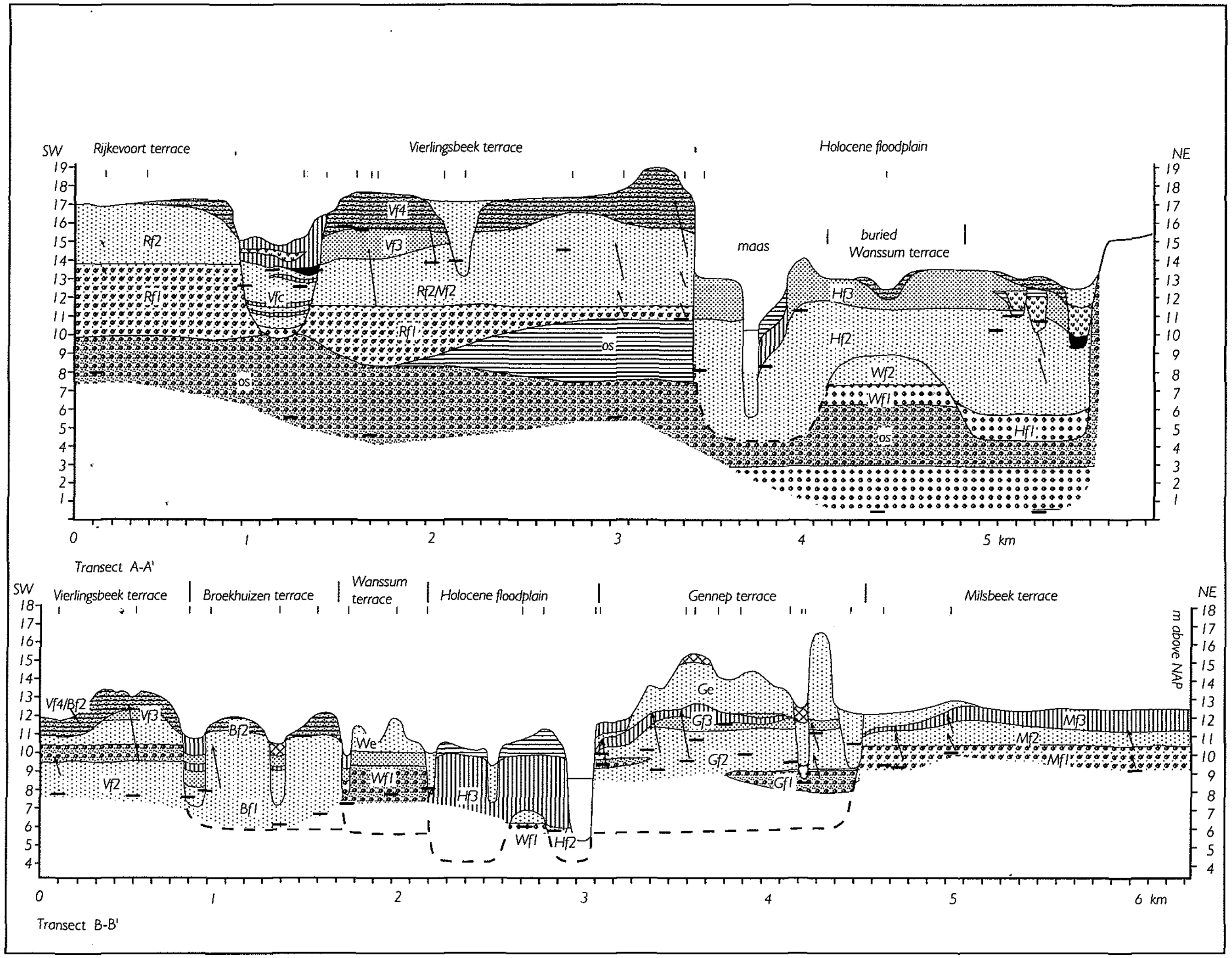


Figure 2.6 Boring transect A-A' and B-B'.

morphology of this terrace is a combination of channel scars of the former braided river system and younger low-sinuosity channels with immature levees. The mean sinuosity is 1.09, which is indicative of a braided river system (Leopold et al. 1964), but the braiding parameter is very low to zero. According to Rust (1978a), this system would be a low-sinuosity single channel system; in this case several low - sinuosity single channels lie on one large floodplain. Transect C-C' (Fig. 2.5) shows the flat nature of the Rijkevoort terrace and the slightly undulating topography of the Vierlingsbeek terrace, resulting from the presence of levees (Vf4, Fig. 2.5), small Late Glacial dunes (Ve, Fig. 2.5) and the slightly incised channels.

Lithologically the terrace sediments consist of sands and gravels in which several facies are distinguished (Fig. 2.5 and 2.6). The lower facies, Rf1, is characterized by an alternation of poorly sorted medium to coarse sand and gravel. Fining-up sequences are not distinct and lithological changes are abrupt. The overlying facies, Rf2/Vf2 and Vf3, consist of fine to medium gravelly sand with up to 2,5 m thick fining-up sequences and gradual lithological changes. The lowermost facies resembles continued deposition by the braided system that also formed the Rijkevoort terrace. This braided system evolved gradually into the transitional system of the Vierlingsbeek terrace and facies Vf2 and Vf3 were deposited; silty, fine-grained sands (Vf4) formed levees. On the terrace surface small dunes (Ve) developed and are characterized by silty medium sand.

The morphology of the Gennep terrace (Niers river) resembles the Vierlingsbeek terrace, although the Gennep terrace is partly covered with aeolian deposits (Fig. 2.2). The gradients of both terraces are comparable (Fig. 2.7), as are the sediments, with coarse, gravelly sands (Gf1, Fig. 2.6) changing upwards into coarse to medium (Gf2, Fig. 2.6) and fine sands (Gf3, Fig. 2.6).

2.4.4 Changes of the Vierlingsbeek terrace upstream of Boxmeer

The boundary between the Rijkevoort and Vierlingsbeek terrace is mostly indistinct, except upstream from Boxmeer where a small escarpment appears between the two terraces. The morphology of the channel scars on the Vierlingsbeek terrace also changes in that area. South of Boxmeer channels are incised up to 6.5 - 7 m (transect A-A' Fig. 2.6) and they are larger and more curved, while the direction of the gullies changes from roughly southeast-northwest (downstream of Boxmeer) to south-north (upstream). The change in depth of incision and in flow direction can be explained in two ways.

A first factor to be considered is the reaction of the Maas to tectonic movements. The longitudinal gradient of the Rijkevoort and Vierlingsbeek terraces between Boxmeer and Wanssum is relatively steep: 35 -41 cm/km (Fig. 2.7). Both up- and downstream the gradients are shallower;,downstream, values of 27.5 cm/km (Berendsen et al. 1995) are observed, and It is likely that the gradient between Boxmeer and Wanssum is influenced by tectonic movements. The Maas flows east of the Peelhorst through the Venlo Graben in this area (Fig. 2.1) and apparently, differential movements of the Venlo Graben occurred during the Late Glacial. Upstream, tilting resulted in a low gradient while downstream subsidence occurred whereby a scarp was formed in the valley. The Maas compensated by incising upstream and depositing sediment downstream of this scarp, which explains the deep incision at Holthees and the scar between the Rijkevoort and Vierlingsbeek terraces in that area.

Another factor that might explain the change in channel morphology north of Boxmeer is the confluence of the River Maas and River Niers, which was a larger river during the Late Glacial than nowadays. According to Zonneveld (1974) the Niers served as one of three branches of the Rhine during the Pleniglacial but was abandoned by the Rhine at the beginning of the Late Glacial. At the confluence large amounts of water and sediment were supplied to the Maas that changed the direction of the channels from south - north to southeast - northwest. It is reasonable to assume that increased sediment input, caused by an higher sediment - water ratio of the Niers, resulted in diminished incision and a more braided river pattern. Such changes, particularly during the unstable conditions of a transitional phase, could cause a considerable effect.

2.4.5 The high-sinuosity meandering river of the Broekhuizen terrace

This terrace is the most easily recognized terrace level in the valley by its large, highly sinuous meander scars, the well-developed pointbars and levees. The preserved pointbars are characterized by fining-up sequences up to 3 m thick, coarse, poorly sorted gravelly sands in the lower pointbar sediments that gradually change upwards into fine-grained, rather well sorted and silty sands at the top of a fining-up sequence. At Beugen (Fig. 2.2) a meander scar, incised up to 5.5 m, shows an asymmetric cross-section (Vandenberghe et al. 1994, Kasse et al. 1995a). Infilling of the gullies consists of peat, gyttja and some sand and is up to 3 m thick. The type of river responsible for these channel scars is a meandering system that incised in the Vierlingsbeek terrace. The areal extent of this terrace is limited because most of the terrace surface was eroded during the Younger Dryas and Holocene. At Beugen a meander neck cut off shows the effect of lateral migration of the meandering river.

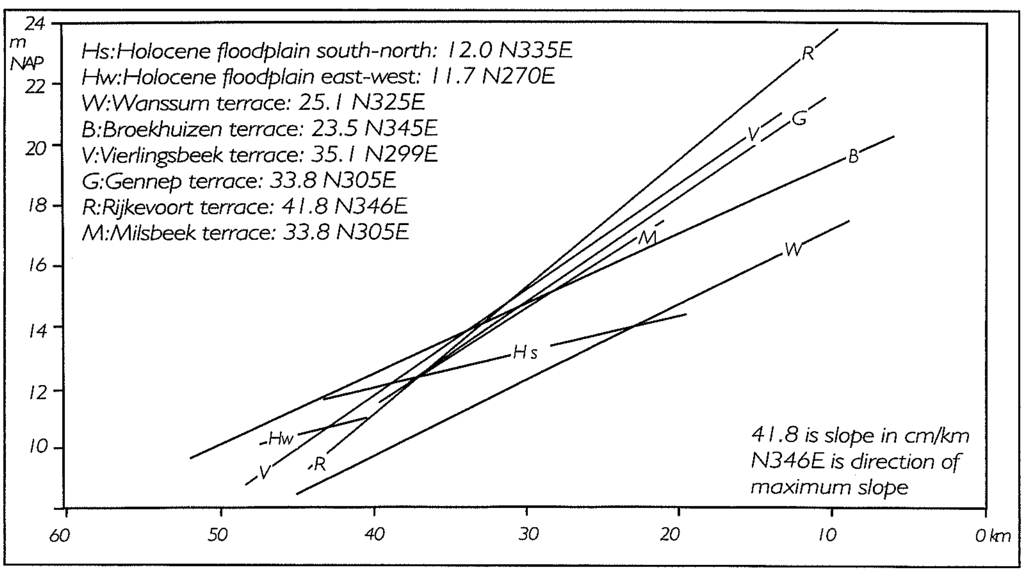


Figure 2.7 Gradient lines of Maas and Niers terraces.

2.4.6 The braiding river system of the Wanssum terrace

The top of the Wanssum terrace is 4 - 8 m lower than that of the Broekhuizen terrace and both the terrace edges and channel scars on the surface are straight (Kasse 1995 b). The sediments of this terrace are coarse grained (850-1400 mu), poorly sorted and gravelly (Wf1, Fig. 2.5), with mostly fining-up sequences of 1 - 2 m, the upper part of which consist of well sorted, fine sands (Wf2, Fig. 2.5). Lithological changes are abrupt, both laterally and vertically. On the east side of the Maas large parabolic dunes of medium grained sand of the same age occur that cover older terrace levels (Fig. 2.2).

The straight morphology of the floodplain and channel scars, combined with the coarse material, poor sorting, short fining-up sequences and abrupt lithological changes indicate deposition by a braided river system. The Younger Dryas palaeo-floodplain has a limited extent because large parts of the Wanssum terrace were eroded in the Holocene. Downstream of Boxmeer, the terrace is also covered by Holocene fine-grained deposits (Hf3, Fig.2.5) and therefore it is not easily distinguished.

2.4.7 The low-sinuosity meandering river of the Holocene floodplain

The Holocene floodplain is a rather narrow, straight-edged plain in which the present Maas meanders have a low sinuosity. In this plain, south of Wanssum, a distinction is made between an older somewhat higher level and the lower-lying active floodplain (Kasse et al. 1995a), which is impossible north of Wanssum because the older level is covered by younger sediments. In drillings however, a distinction between different channel systems is possible, and in transect A-A' (Fig. 2.6) two, up to 8 m deep incisions can be seen. The lowermost sediments are coarse (Hf1, Fig. 2.6), but most of the floodplain sediments consist of fine sediments: fine-grained and sometimes clayey sand, silt and some organic materials (Hf2, Hf3 Fig. 2.5 -2.6) that were deposited in levee and overbank environments.

2.5 AGE AND CORRELATION OF THE TERRACES

Table 2.1 shows the correlation of this local terrace subdivision with terrace divisions proposed both upstream (Kasse et al. 1995a) and downstream (Berendsen et al. 1995). The correlation is complicated by the fact that the upstream division was achieved mostly using morphology, an approach which is not possible for the area downstream of the terrace intersection where the division is based on lithology and subsurface morphology. The age estimation of the terraces is mostly based on stratigraphical position and correlation with dated terrace surfaces elsewhere. In this area no radiocarbon dating was performed since most sediments were sterile and the scarce organic sediments found were too young and originated from rather recent floodings in the Holocene.

On the Rijkevoort terrace, Kasse et al. (1995a) were able to distinguish a level 1, covered with aeolian sands, and a level 2 lacking coversands. The Rijkevoort terrace has also been found in the Land van Maas en Waal. Radiocarbon dates from the terrace sediments are not available because of the sterile nature of the sediments, but the age is most probably Weichselian late Pleniglacial as the aeolian sands on the terrace at Rijkevoort resemble the Pleniglacial aeolian coversand facies of the Netherlands. The aeolian sediments on the Rijkevoort terrace correlate with the sediments described upstream near Grubbenvorst (Mol et al. 1993) which were also thought to be of late Pleniglacial age.

The transitional system of the Vierlingsbeek terrace is both recognized by Kasse et al. (1995a) as "level 3", and by Berendsen et al.(1995) as "generation 2" (Table 2.1), the latter comprising both the Rijkevoort and Vierlingsbeek terraces. The morphological differences between the two terrace surfaces are already small downstream of Boxmeer and no distinction could be made by Berendsen et al. in the Land van Maas en Waal. The generation 2 river system is described as braided though some major curved channels have been recognized (Berendsen et al. 1995, Makaske and Nap 1995). As the Vierlingsbeek terrace is younger than the late Pleniglacial Rijkevoort terrace and older than the Broekhuizen terrace of Allerød age, this terrace is most likely of Bølling age, which is confirmed by dates of organic deposits in abandoned channels of the Vierlingsbeek terrace. Channels south of Nijmegen; 12.780 ± 110, 12.640 ± 310 BP (Teunissen and De Man 1981), and 12.520 ± 180 BP (Teunissen 1990) and in de Hamert (12.760 ± 150 BP, Teunissen 1983) were dated. Locally, an abandoned channel on the older, Rijkevoort terrace, that still functioned during the Bølling, has been dated, such as at Bosscherheide (12.110 ± 70 BP, Bohncke et al. 1993).

The Broekhuizen terrace can be recognized throughout the valley and organic infillings of abandoned gullies show pollen assemblages of the Allerød and Younger Dryas periods. The meander at Beugen was active in the early Allerød (Kasse et al. 1995a), but most dates of organic infillings are of Younger Dryas age (Berendsen et al. 1995), indicating that the meandering system was active until the very end of the Allerød. The Wanssum terrace correlates with level 5 of Kasse et al. (1995a), generation 4 of Berendsen et al. (1995) and with "terrace X" of Pons (1957). The Younger Dryas age of this system is well established by pollen analyses and C14 dates by various authors (Berendsen 1988, Westerhoff and Broertjes 1990, Kasse et al. 1992, Bohncke et al. 1993). The Younger Dryas dune deposits on the east side of the Maas valley, overlying older terrace surfaces there, are dated from the later part of the Younger Dryas at Bosscherheide (Bohncke et al. 1993).

The River Maas probably incised the Younger Dryas floodplain at the very beginning of the Holocene, as parts of the floodplain were already abandoned in the early Boreal and organic materials accumulated (Kasse et al. 1992). The shape of the Holocene floodplain is largely inherited from the former period and the Maas was confined to a relatively narrow floodplain for the whole Holocene. Downstream of Nijmegen the Maas occupied a more extensive floodplain by occupying several channel systems (Berendsen et al. 1995).

2.6 SEDIMENT PETROLOGY

Heavy mineral analysis is used to distinguish Late Glacial from older deposits, to determine the base of the Late Glacial sediments, to differentiate between the Late Glacial terraces, to differentiate Niers from Maas sediments and to determine the origin of the various aeolian deposits (Fig. 2.8). The Saalian and Weichselian sediments in the Maas valley belong to the Kreftenheye Formation, which consists of both Maas and Rhine sediments (Verbraeck 1984). During the late Pleniglacial and Late Glacial these sediments were reworked and mixed with fresh Maas sediments and as a result a mixed heavy mineral assemblage was formed. Typical Maas minerals including tourmaline, stable minerals (zircon, rutile, anatase) and Rhine-derived minerals like augite (volcanic) and hornblende are present.

The composition of the Middle Pleistocene sediments on the Peelhorst (Overloon terrace) differs from that of the Weichselian Maas terraces by lower contents of garnet, epidote and hornblende and a higher amount of tourmaline (Fig. 2.8). According to Zonneveld (1947) these sediments belong to the 'Veghel Zone' of Saalian age. Downstream of Boxmeer, the confluence of the River Niers provided fresh Rhine- derived minerals, which show an assemblage that is distinctly different from the Maas sediments. A high frequency of volcanic minerals (augite, enstatite, hypersthene, titanite, anthophyllite, glaucophane, olivine) and hornblende, and a low quantity of tourmaline indicates a stronger Rhine influence; the increased frequencies of volcanic minerals being derived from the Eiffel (Zonneveld 1974). The confluence of the River Maas and River Niers is typified by their intermediate mineral composition downstream of the confluence (Fig. 2.8). The relative importance of the Niers sediment in the Late Glacial terraces is great in the confluence area.

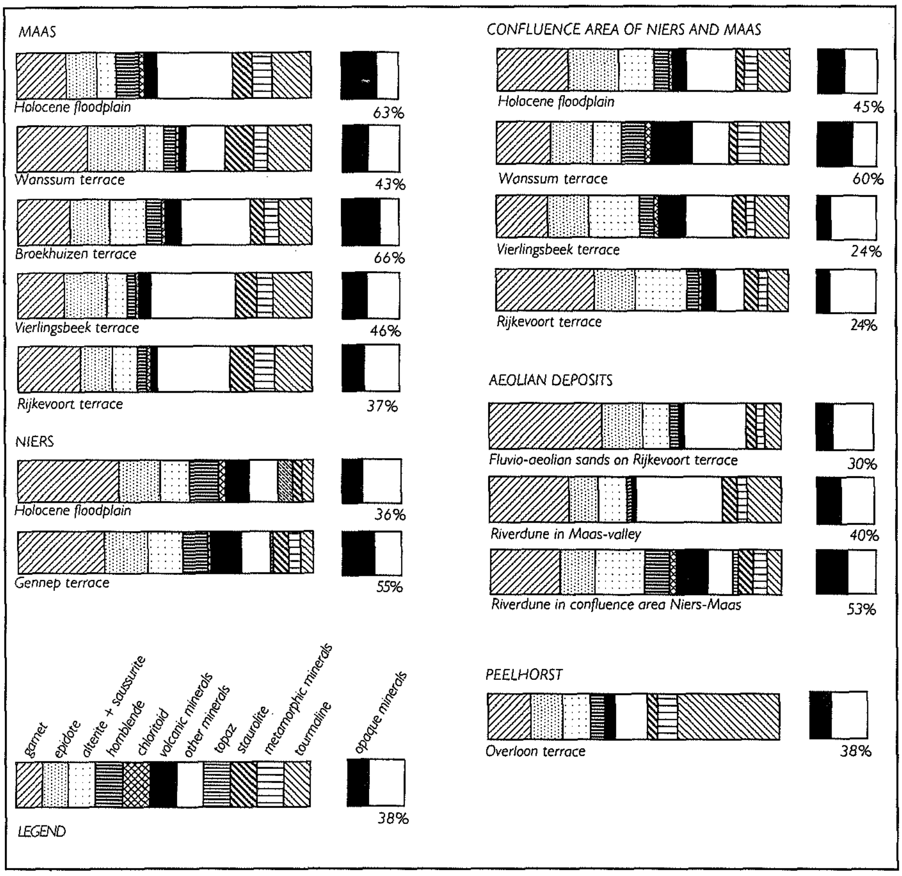


Figure 2.8 Average heavy mineral contents of Maas and Niers terraces and from aeolian deposits.

An ambiguous trend is seen in the sequence of the Late Glacial Maas terraces, from the Rijkevoort terrace to the Holocene floodplain, with the exception of the Wanssum terrace . In these terrace sediments the content of garnet diminishes from 22 to 16 % and the amount of opaque minerals increases from 37 to 63 %. Towards the younger terraces, the Maas component becomes slightly more dominant in the sediment, a trend that is best illustrated by an increase in opaque minerals, reflecting decreased reworking of older sediments. The Wanssum terrace sediment does not fit in this trend as the minerals show a higher garnet content (24 %), a high epidote content (19 %), a low amount of other minerals that are predominantly stable minerals like zircon and rutile (14%) and a distinctly lower amount of opaque minerals (43 %). The mineral assemblage of this terrace sediment shows a stronger Rhine influence which is related to the deep incision and reworking of older material during the Younger Dryas.

The Younger Dryas river dune sands were derived from the Wanssum palaeo-floodplain because the heavy mineral composition in the aeolian sands strongly resembles the composition of the Wanssum terrace sediment. Upstream of the confluence with the River Niers the Younger Dryas fluvial and aeolian sediments show 12-15 % tourmaline and 40-43 % opaque minerals, while downstream of the confluence they both show 5-9 % tourmaline and 53-60 % opaque minerals and a high amount of volcanic minerals. The source area of the aeolian deposits is therefore most likely the braidplain. In such a river system the floodplain is inactive over large periods of the year and it can therefore easily supply sediments for aeolian transportation.

2.7 PHASES OF INCISION AND ACCUMULATION DURING THE LATE PLENIGLACIAL AND THE LATE GLACIAL

Phases of incision and accumulation have been reconstructed from borehole transects (Fig. 2.5 and 2.6) and are shown in Fig. 2.9. During the late Pleniglacial, accumulation became dominant and Rf1 and Rf2 (Fig. 2.5 and 2.6) were deposited. As the system became less active finer grained sediments were deposited and, locally, aeolian sands accumulated. The thickness of the Rf1 and Rf2 deposits is about 7 to 9 m (Fig. 2.6 transect A-A').

At the beginning of the Late Glacial some gullies of the existing braided system started to incise up to 3 m (Fig. 2.5) while at the same time levee and overbank deposits (Vf4) accumulated so on a floodplain scale the net result is neither incision nor accumulation. The major channels continued to incise while the minor ones were abandoned until finally one large channel remained active in the Allerød. This large meandering channel incised up to 6 m and shifted laterally whereby pointbar sediments (Bf1) accumulated to the same height or even slightly higher than the surface of the former terraces. Overbank and levee deposits (Bf2) were deposited on top of Bf1, but also on the older Vierlingsbeek terrace (Fig. 2.6, transect B-B', Vf4/Bf2). Makaske and Nap (1995) also recognized incision, lateral erosion and accretion by a large meandering channel downstream in 'the Land van Maas en Waal' (Fig. 2.1). The accompanying pointbar formation ceased during the Allerød or early Younger Dryas.

An abrupt incision took place at the end of the Allerød and beginning of the Younger Dryas on a floodplain scale when channels incised at least 4 m deeper than the base of the Allerød channel (transect A-A' Fig. 2.6) and 8 to 10 m sediment of former terraces was eroded in the Younger Dryas floodplain. The scale of incision, over the total width of the floodplain, is large in contrast with the incision in the former period, where a single channel moved laterally in the floodplain, incising and depositing sediments at the same time. The lateral dimensions of the Younger Dryas floodplain are comparable to the present Holocene floodplain. After or during this major incision coarse material was deposited (Wf1), covered mostly by coarse sandy deposits (Wf2) but the net accumulation was not sufficient to fill the entire incision. The top of the Wanssum terrace is therefore 3 to 4 m lower than the surfaces of former terraces (Fig. 2.5-2.6). At the transition from the Younger Dryas to the beginning of the Holocene, channels again incised into the Younger Dryas floodplain (transect A-A', Fig. 2.6) and the base of the Holocene channels is at least 1 to 2 m deeper than the base of the Younger Dryas sediments. Some channels were filled with coarse material, (Hf1), while others remained active. For the largest part of the Holocene one channel shifted in the floodplain, while depositing fine grained sediments (Hf2 and Hf3).

2.8 IMPACT OF TECTONIC ACTIVITY ON THE FLUVIAL EVOLUTION

Longitudinal gradient lines of Late Glacial terraces of the Maas (Fig. 2.7) provided some insight in the Late Glacial tectonic activity (Huisink 1998). Gradient lines of both the Rijkevoort terrace, (41,8 cm/km), and the Vierlingsbeek terrace, (35,1 cm/km) are steep, when compared with downstream values of respectively 25,3 cm/km and 27,5 cm/km (Berendsen et al. 1995). Gradient lines of the younger Broekhuizen terrace, 23,5 cm/km, and Wanssum terrace, 25,1 cm/km, are much less steep than those of the Rijkevoort and Vierlingsbeek terraces in the same area and compare better with downstream values of 22 cm/km (Verbraeck 1990) and 25,4 cm/km (Berendsen et al. 1995) respectively. It appears that the Rijkevoort and Vierlingsbeek terraces have been subjected to differential movements in the Venlo Graben. A more detailed study of the gradient line on the Vierlingsbeek terrace by Huisink (1998) revealed a difference between the downstream part of the terrace and the upstream part. The downstream part had a lower gradient value compared to the upstream part but this had no consequence on the channel morphology. Furthermore, it appeared that river patterns and terraces are traceable throughout the Maas valley, regardless of areas of lower or higher gradient values. Tectonic activity, therefore, had a minor influence during the formation of terraces, but it locally influenced the depth of the incision (Huisink 1998). In Fig. 2.7 the steady decrease of the gradient from the late Pleniglacial to the Holocene is demonstrated. The gradient of the Allerød meandering river of 23,5 cm/km is quite similar to the gradient of the Younger Dryas braided river (gradient of 25,1 cm/km) in contrast to their fluvial styles. The fluvial style of the Maas is thus not noticeably changed by the slope of its floodplain.

The influence of tectonic activity on the Maas valley is twofold. i. The long term gradual uplift of the hinterland resulted in a tendency of the river to erode which is a long term process and resulted in the preservation of older fluvial sediments in terraces. The formation of at least three terraces in 3000 years during the Late Glacial, however, cannot be attributed to this long term gradual process. ii. Locally, tectonic movements resulted in deeper incision, which affected a small area but the steepness of terrace slopes did not noticeably change channel morphology (Huisink 1998).

2.9 IMPACT OF CLIMATE ON THE FLUVIAL EVOLUTION

2.9.1 Climate reconstruction, vegetation and river development

A very cold and dry climate with permafrost conditions prevailed during the late Pleniglacial (Vandenberghe 1991) with sparse, or even no vegetation (Hoek 1997a). The combination of periodically high discharges, caused by snow melt, and a high sediment supply, caused by the absence of vegetation, resulted in a highly energetic braided river (Fig. 2.4). Towards the end of the Pleniglacial, mostly aeolian sediments were deposited in the Netherlands, thus indicating drier conditions. Permafrost conditions probably no longer occurred (Van Huissteden and Vandenberghe 1988). During this time the river system became active in the eastern part of the floodplain while coversands were deposited over the inactive western part. A less energetic braided river system evolved with somewhat finer material and smaller gully dimensions (Fig. 2.4).

At the onset of the Late Glacial the temperature rose and wetter conditions occurred as is shown by high lake levels (Bohncke and Wijmstra 1988). A herbaceous vegetation with dwarf bushes colonized the surface

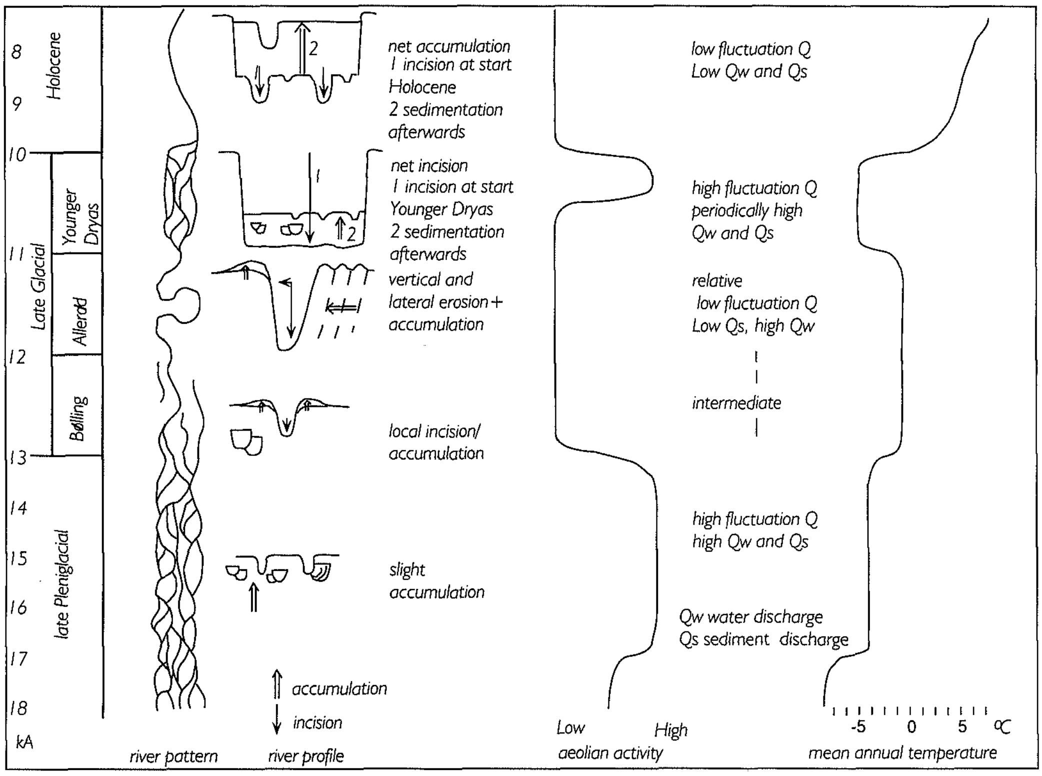


Figure 2.9 Summarized synthesis of fluvial development of the Maas.

in the early Bølling, resulting in a denser plant cover on the surface and a diminished sediment load. The landscape, soil and vegetation indicated by the Usselo site had similar characteristics to the modern calcareous coastal sand dunes in Denmark, The Netherlands and Belgium (Wiegers and Van Geel 1983), which was a sandy habitat with a low vegetation cover during this period. In Opgrimbie, Belgium, a relatively open landscape with birch was also recognized by Paulissen and Munaut (1969). Although the surface was colonized by vegetation, the protective plant cover was not dense and continuous , so the soil was exposed to erosion by rain and wind (Wiegers and Van Geel 1983). There was, however, more protection against erosion than in the previous period which resulted, together with a more regular discharge regime, in a river pattern change. Minor shallow channels became abandoned, or functioned only at very high water levels (Pons 1957), while larger channels began to incise. This was the transitional stage, from a braided to a meandering river. Although grasses and herbs colonized the surface rather fast, it was not until the middle to late Bølling that birch colonized the surface (Paus 1995) and an open tundra to boreal birch forest was established (Bohncke and Wijmstra 1988). The soil became increasingly protected against erosion and incision of the major channels proceeded. The vegetation developed into a more dense vegetation cover of open birch woods at the end of the Bølling (Van Geel et al. 1989), containing a small percentage of pines in Belgium (Paulissen and Munaut 1969) and ultimately, only one major, highly sinuous meandering, 5 to 6 m incised channel developed in the Allerød.

At the beginning of the Younger Dryas the climate was considerably wetter and cooler (Bohncke et al. 1987, 1993). Discontinuous permafrost was present, as is indicated by the presence of permafrost-related periglacial phenomena such as cryoturbated silts and indistinct wedge forms in Bosscherheide (Fig. 2.4). The vegetation cover was initially still intact, so bank and soil stability were maintained and sediment supply was restricted. Since the discharge regime changed immediately and became greater and more irregular, an incision occurred, probably in a very short period of time. The transition from Allerød to Younger Dryas was recorded at Usselo as peat vegetation, which changed rapidly into a sandy deposit. This deposition of sand indicated erosion and deposition related to an incomplete vegetation cover (Van Geel et al. 1989). As soon as the forest vegetation was partially destroyed and changed to a more open shrub vegetation with heather (Bohncke and Wijmstra 1988), the sediment load increased and the braided river system partially filled the incised floodplain. From 10.550 BP the climate became drier and warmer (Bohncke et al. 1993). Around this time the large parabolic dunes developed on the east side of the Maas. The sediment of the dunes originated from the Younger Dryas floodplain. This illustrates the dynamics of the braided river system, in which parts of the floodplain were periodically inactive and formed the source area for aeolian transport.

Temperatures rose at the onset of the Holocene and closed birch woods quickly became re-established on the river floodplain (Hoek 1997a). The river banks were stabilized and relative sediment load was low. The Maas reacted by incising and changing its pattern again to a meandering river.

2.9.2 Correlation with other Northwest European and Central European lowland rivers

Changes as witnessed in the Maas valley can be compared with other European lowland rivers valleys where Late Glacial climatic changes had comparable impact on the rivers although they have different catchments and tectonic settings. Evidence for Late Glacial river pattern changes in the Ardennes, the hinterland of the Maas, is scarce since erosion dominated here. Paulissen (1973) recognizes only one Late Glacial terrace level, which is formed by sediments of a Younger Dryas braided river, and a meandering river in the Holocene in the Maas valley north of Maastricht (Fig. 2.1). More complete Late Glacial fluvial records are found in the western part of Belgium, in the middle and lower River Scheldt valley. Here, the transition from a braided to meandering pattern at the onset of the Late Glacial is described by Kiden (1991) but a change in pattern during the Younger Dryas is not observed.

The warming of the Late Glacial is witnessed in lowland Britain by a change to a more regular river regime (Rose 1995). A meandering phase in the Allerød is not distinguished but the cooling of the Younger Dryas resulted also in an incision, up to 9.5 m. Aggradation followed the incision in the later Younger Dryas, which correlates with the Maas. The fluvial changes are explained by snow melt and sediment yield that are related to soil protection by vegetation. Collins et al. (1996), described in a recent study a transition from a braided river to a meandering state in the Allerød, based on pollen analysis and C14 dates. A highly energetic river regime was active during the Younger Dryas. The succession described here closely resembles the fluvial development of the Maas.

The transition from a braided river into a meandering one via a transitional phase at the onset of the Late Glacial is also seen in river valleys in Germany (Schirmer 1983, Lipps and Caspers 1990). A deep incision at the Allerød - Younger Dryas transition occurred in the Ems (10 m) and Niederrhein (14 m) valleys (Klostermann 1995). The incision in the Niederrhein valley in Nordrhein Westfalen is entirely filled in by deposition of sediments in the Younger Dryas period, in contrast to the Ems valley where the amount of sedimentation afterwards was not sufficient to fill the entire incision and resulted in a lowering of the terrace surface of 5 m. Again, these river reconstructions match the reconstruction of the Maas quite well.

An extensive comparison of the Maas with the Warta river in Poland has been made by Vandenberghe et al. (1994). Both rivers have experienced similar changes except for the Younger Dryas period, when the Maas changed into a braided river while the Warta remained a meandering river. In a number of Polish river valleys two phases of meandering channels are found; a Late Glacial phase of large, highly sinuous meandering channels and a Holocene phase of smaller sinuous channels (Szumanski 1983, Kozarski 1983). In the upper Vistula valley, remnants of a braided river system in the Younger Dryas (Kalicki and Zernickaya 1995) are found, though not throughout the whole valley section (Starkel 1995). Both the Warta and Vistula adapted a meandering form in the Holocene and reacted to the Late Glacial climatic changes in a comparable manner to the Maas.

In the Somme valley, Northern France, the late Pleniglacial transition to the Late Glacial is also characterized by a change from a braided river into a multi-channel transitional system, in which organic deposits of Bølling and Allerød ages are found. Infilling mostly occurred during the Younger Dryas, while the second important incision took place at the onset of the Holocene. The development of the Somme valley is again quite similar to that of the Maas, except for the major incision phase around 11.000 BP in the Maas valley.

2.9.3 Model for fluvial changes related to climatic changes

Fluvial changes in the Maas valley after the last glacial maximum can, almost certainly, be linked with climate-related changes in sediment and water supply. Vandenberghe (1993) proposed a non-linear model for fluvial changes related to climatic changes, stating in short that incision takes place at the start of a cold period, linked to a delayed disappearance of vegetation compared to temperature decline. Incision also takes place at the transition from a cold to warm period whereby incision is induced by a delay in full forest development. Such phases of incision were also found in this study, namely at the transition from the colder Pleniglacial to the warmer Late Glacial, at the transition from the warmer Allerød to the colder Younger Dryas and finally the transition from the colder Younger Dryas to the warmer Holocene. The time lag between fluvial and climate changes can be explained in two ways. i. Vandenberghe (1995) concluded that the delayed response of the Maas around 12-13 ka represents an intrinsic evolution over some 1300 years. ii. Another possibility is that this river pattern change does not necessarily represent an intrinsic evolution but is the direct response of the river to changing discharge characteristics and sediment load.

Only one gradual change in pattern is seen in the Bølling, where the braided river changed via a transitional system to a meandering river, and two abrupt changes (Allerød - Younger Dryas, Younger Dryas - Holocene) in channel morphology. When the transitional system is regarded as intrinsic evolution of the river, the question may arise why this intrinsic evolution occurred only at the onset of the Late Glacial and not at the onset of the Holocene or at the transition from the Allerød to Younger Dryas as well. Apparently the intrinsic evolution occurred favorably under changes in climate from cold to warm and not in the reverse situation. According to Bohncke et al. (1987) the incision began at the very beginning of the Late Glacial as soon as vegetation colonized the surface, with the fluvial pattern changing gradually afterwards, as a result of intrinsic evolution. The surface was still rather open in the Bølling, however, and it was not until the start of the Allerød that a more closed vegetation cover had developed, which could point to a more gradual decrease of erosion than envisaged by Bohncke et al (1987). The change from almost bare vegetation via herbs and shrubs to forest vegetation influences the erosion rates greatly, as Kirkby (1980) showed that slope erosion by water increasingly diminishes when comparing bare land, shrub, grassland and forest. Forest is the most erosion free environment due to the high infiltration capacity and percolation. Trees (especially needle-leaf trees, like pines) produce organic layers with high infiltration capacity (Trimble 1988), suggesting a considerable decrease in slope erosion rates during the Bølling period. In this hypothesis, this should have resulted in an increasing incision of the larger channels so that the river system reacted in a direct manner to changing conditions, rather than in an intrinsic manner.

Another way of explaining the fluvial changes of the Maas during the Late Glacial is to consider a direct response of the river to changing vegetation, discharge characteristics and sediment load. Hence, the sudden changes in fluvial pattern at the transition of the Allerød to the Younger Dryas and from the Younger Dryas to the Holocene can be explained by the rapid change in vegetation and discharge characteristics, while the gradual change at the start of the Late Glacial is related to gradual changes in sediment supply, combined with a gradually changed discharge regime. The re-establishment of closed vegetation at the onset of the Holocene was relatively fast in contrast to the onset of the Late Glacial, where a time lag of up to a 1000 years occurred in Northwest Europe between the warming of the Bølling period and the first birch forest development (Paus 1995). Vegetation cover is an important factor in determining the river pattern since it controlled the amount of erodible sediment.

Another important factor, apart from the vegetation cover, in explaining the river pattern changes is changing discharge characteristics, which is applied in both explanations mentioned above. The mean annual temperatures rose quickly at the start of the Holocene, just as at the start of the Bølling. The winter temperatures remained low in the Bølling (Bohncke and Vandenberghe 1991) however, so that snowfall in winter was high and the soil was frozen for major parts of the year. Discharges in the Bølling were, although more regular than during the late Pleniglacial, still rather irregular and depended for a large part on snowmelt, while in the early Holocene, discharge was determined less by snowmelt and therefore more regularly distributed throughout the year. In addition, water storage in the soil was relatively low when compared to the early Holocene. The greater storage of water in the soil and less irregular discharge, combined with the low slope erosion rates which are linked to the rapid vegetation change, favoured the accelerated change to a meandering river in the Holocene.

The river pattern changes of the Maas can thus be explained in two ways: either as a direct response to climatically-derived extrinsic factors or as an intrinsic evolution. The climate-related factors (i.e. changes in vegetation type and coverage, discharge characteristics and soil water storage capacity) influence one another so it is difficult, if not impossible to determine which factor is the most important, however, different rates of vegetation development are a significant determinant in the morphological changes observed in the Maas valley in both hypotheses.

2.10 CONCLUSIONS

It is shown that Late Glacial morphological and sedimentological changes of the Maas occurred throughout a large part of the valley, regardless of local steeper gradients that are related to local tectonic movements. The influence of tectonic activity is twofold: i. the uplift of the hinterland resulted in a tendency to erode and preserve older terraces. ii. locally deeper incision of channels occurred related to steeper gradients. Changes of the Maas river patterns can be explained by changes in water storage capacity of the soil (permafrost), water discharge (relative importance of snow melt water) and, most importantly, sediment supply related to vegetation cover, which are all climate-induced factors. Major incisions took place at transitions from a warm to cold period (Allerød to Younger Dryas) and cold to warm period (Pleniglacial to Bølling, Younger Dryas to Holocene). A slow change of river pattern from a braided river system to a meandering river system with an intermediate transitional system is observed at the beginning of the Late Glacial, which can be explained in two ways, either an intrinsic evolution of the Maas or the direct response of the Maas to the slow establishment of a closed vegetation cover. In both cases it can be stated that the rate of river pattern change is dependant on the rate of adaption of vegetation to changing climatic conditions. At the start of the Holocene a closed vegetation recovered much faster, which resulted in an abrupt change from a braided to a meandering river. The non-linear behaviour of the Maas to climatic changes, proposed previously (Vandenberghe 1993), is confirmed by this study and other studies of Northwest and Central European lowland rivers support the conclusion that river pattern changes are triggered by climatic changes.

3 Tectonic versus climatic controls on the River Maas dynamics during the Lateglacial

by M. Huisink. Published in Benito, G., Baker, V.R. and Gregory, K.J. (eds.) : Palaeohydrology and Environmental Change, p 99-109 in 1998.

3.1 INTRODUCTION

The Maas (Meuse) is a rainfed river which drains an area of 33.000 km^2. It flows from France through Belgium, and enters The Netherlands from the south (Fig. 3.1). There it flows through the southern part of the North Sea Basin which is filled with Quaternary sediments. An important part of the headwaters of the Maas is formed by the relative high Ardennes, that are the result of long term tectonic uplift. The Maas crosses a part of the Roer Valley rift system, namely the South Limburg Block, the Roer Valley Graben, The Peel Block (Peelhorst), and it finally flows through part of the Venlo Graben, where the study area is situated south of Nijmegen (Fig. 3.1). Since the hinterland has been subjected to uplift the Maas had a tendency to erode and formed a series of terraces. During the Lateglacial (13 ka - 10 ka) the Maas incised several times and deposited sediments afterwards in an increasingly narrowing floodplain. Five terraces are recognized from the Weichselian late Pleniglacial (oxygen isotope stage 2) to the Holocene.

The alternation of erosion and deposition resulted in a complex morphology, which is enhanced by the fact that incision is often equally balanced by deposition. During the Lateglacial the sealevel was some 100 m lower than present (Jelgersma 1966), but because the sealevel rose considerably during the Holocene, sedimentation of holocene sediments became dominant until roughly this area. Upstream erosion prevailed, while downstream younger sediments are deposited on top of older terraces. Reconstructions of the Maas fluvial activity were made by Pons (1957), Pons and Schelling (1951), Berendsen et al. (1995), Vandenberghe et al. (1994), Kasse et al. (1995a), Bohncke et al. (1993), Van den Berg (1994a, b, 1996). Recently, a morphological and sedimentological study was undertaken to refine the reconstruction of the river evolution (Huisink 1997).

However, the relative importance of either climate-related factors or tectonic-related factors on the river system is still not completely understood. The focus of this paper is therefore to study the impact of both climatic and tectonic controls on changing fluvial patterns and phases of incision and accumulation. More insight into the tectonic control on river patterns is achieved by the construction of longitudinal profiles of each terrace by multiple linear regression analysis. Locally steep terrace slopes are recognized which are explained by local tectonic movements. The effect of these steeper terrace slopes on river morphology and sedimentology will be discussed.

3.2 LATE WEICHSELIAN TERRACE STRATIGRAPHY

The terrace stratigraphy of the Maas valley, south of Nijmegen, has been made by Huisink (1997) and is shown in Fig. 3.2. Detailed sediment descriptions and ages of sediments are provided by Huisink (1997).The oldest investigated Weichselian terrace level is the Rijkevoort terrace of Pleniglacial age. The terrace sediments were deposited by a braided river. Part of the valley was abandoned during the Late Pleniglacial and became covered by 0 to 4 m aeolian sands (so-called coversands).

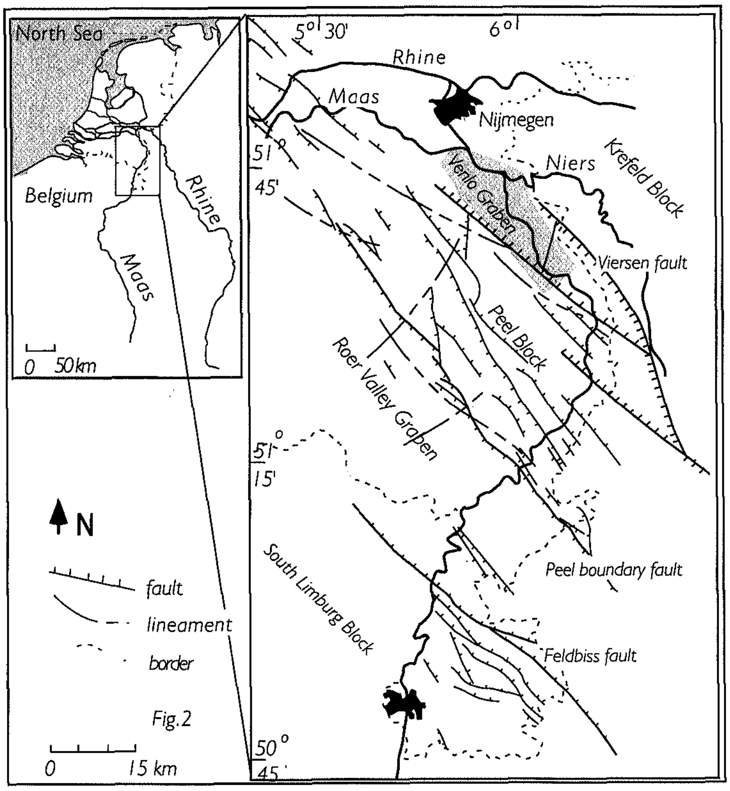


Figure 3.1 Location of the study area and major faults (faults according to Van Montfrans, 1975, lineaments with signs of horizontal motion according to Van den Berg, 1994b).

The Pleniglacial river system changed at the beginning of the Lateglacial (13 ka), when minor channels of the braided system became abandoned and larger ones started to incise up to 3 m. The morphology shows abandoned channel scars of the former braided river and low-sinuosity channels with immature levees of the new river system (Vierlingsbeek terrace). Upstream of Beugen the channel morphology of the Vierlingsbeek terrace changes (Fig. 3.2). At some locations several small, low-sinuosity channels merged in one large, more curved channel (Vandenberghe et al. 1994). This river system represents a transition between the former braided river system of the Rijkevoort terrace and the meandering river of the Broekhuizen terrace. The incision of the major channels continued until finally one, highly sinuous channel remained active in the Allerød (12 - 11 ka, Broekhuizen terrace).

At the onset of the Younger Dryas (11 ka) the river system changed rapidly into braiding (Wanssum terrace). A distinct incision took place whereafter the floodplain was partly filled in by sediment. Aeolian processes became important in the later part of the Younger Dryas when large parabolic dunes were formed on the east banks of the Maas. The sediment for these dunes originated from the Younger Dryas floodplain, as shown by heavy mineral analysis (Huisink 1997). Finally the Maas incised the Younger Dryas floodplain at the onset of the Holocene and changed again into a meandering river. The dimensions of the

Holocene floodplain are largely comparable to those of the Younger Dryas floodplain. The Holocene Maas was a low sinuous river confined to a relatively small floodplain in which the channel shifted. The Holocene floodplain widens downstream of Nijmegen where five Holocene river systems were recognized by Berendsen et al. (1995).

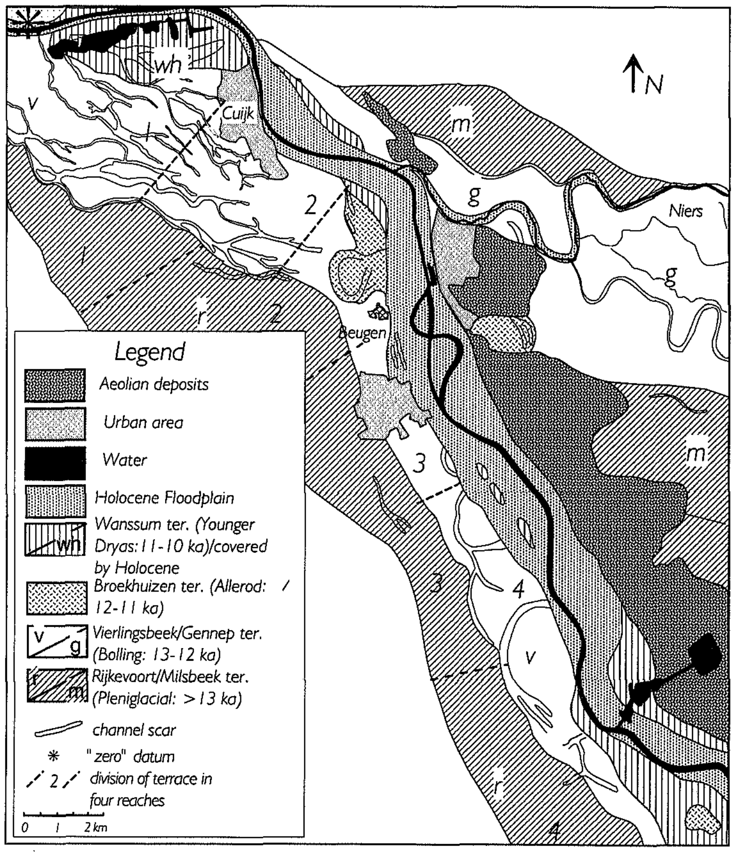


Figure 3.2 Late Weichselian terraces and Holocene floodplain of the Maas.

3.3 CALCULATING LONGITUDINAL PROFILES OF RIVER TERRACES

Longitudinal profiles of the Lateglacial terraces are constructed to improve the terrace stratigraphy and to establish the effects of tectonism on river pattern changes. A reliable reconstruction requires a large number of altitudes and an objective way of plotting. A multiple linear regression (mlr) analysis is used therefore instead of the more traditional perpendicular-projection of ground point data to a hand drawn river axis transect (e.g. Merritts et al. 1994). A mlr-analysis determines the relation between the independent variables (x and y coordinates) and a dependant variable (z or altitude) by fitting a planar surface through the data. The assumption that a regression surface resembles a terrace can be made in this area because the valley has a more or less straight direction. When a large bend occurs in a valley a curved surface would be needed, resulting in a more complicated regression equation. Since the x and y coordinates are equally important a step wise mlr analysis has not been used.

The data points are collected from altitude maps with a scale of 1 to 10,000. Only the highest points on the terrace surfaces are used, mostly the points on top of top of levees or pointbar sediments. Detailed sedimentary and morphological analysis is necessary to exclude aeolian deposits. A minimum of 54 and a maximum of 133 points for each terrace are used in the mlr-analyses. The output of the mlr analysis for each terrace is an equation representing the best fit of a surface through the observed points (Fig. 3.3). This equation consists of a constant and two regression coefficients which are calculated by the method of least squares. These regression coefficients explain the effect of the coordinates on the altitude. The standard error of estimate is a measure for the scatter of the data around the regression line and the R-squared, goodness of fit, represents the degree of association between the measured and estimated altitudes. The slope of the surface, calculated by the mlr analysis, is the steepest longitudinal terrace profile and is determined by taking the root of the combined squared regression coefficients (x coeff.^2 + y coeff.^2= surface slope^2).

The surface, calculated by the mlr-analysis, is a three dimensional feature. In order to visualize the surface slopes in two dimensions, the altitude has to be plotted against one axis. This can be accomplished by rotation of the x-y coordinate configuration in the direction of the steepest surface slope. To obtain this rotation, the most distal point in the area is taken as "zero" point (Fig. 3.2). A line is calculated through this point, the so called zero-line, which has a direction perpendicular to the surface slope direction. The coordinates of all points are recalculated in distances from this "zero" line. The distances of points can now be plotted against altitude and a linear graph is constructed which represents the longitudinal terrace profile. The longitudinal profiles of all terraces are plotted in figure 3. 4 for comparison. The Holocene floodplain was divided into two reaches, up- and downstream of Cuijk (Fig. 3.2) because the direction change in the holocene floodplain appears to be too large to consider the whole floodplain as one flat surface.

3.4 RESULTS

The reliability of these reconstructions (Fig. 3.3) is high: the R-squared varies between 0.92 (Allerød terrace) and 0.99 (late Pleniglacial terrace). The Holocene floodplain downstream of Cuijk (Fig. 3.3f) is poorly constructed (R2 = 0.6), therefore only the more reliable reconstruction of the floodplain upstream of Cuijk (Fig. 3.3e) is discussed further. The scatter of points around the regression line is normal for all terraces (68% of the points are within the area of plus or minus the standard error of estimate). Points with a large deviation from the regression line were checked again and eventually corrected if in error.

The oldest terraces are relatively steep, 41.8 cm/km for the Rijkevoort terrace (late Pleniglacial) and 35.1 cm/km for the Vierlingsbeek terrace (Bølling) when compared with the slopes of the Broekhuizen- (Allerød, 23.5 cm/km) and Wanssum terraces (Younger Dryas, 25.1 cm/km). The longitudinal profiles of the Broekhuizen and Wanssum terraces correspond well with downstream values of resp. 22 cm/km (Verbraeck 1990) and 25.4 cm/km (Berendsen et al. 1995). The Rijkevoort and Vierlingsbeek terraces appear to be too steep, as downstream 27,5 cm/km is observed by Berendsen et al. (1995) for their "generation 2 " river system which comprises both the Rijkevoort and Vierlingsbeek terraces (Huisink 1997). Upstream, also a gentler slope value of 22 cm/km for the Rijkevoort terrace (Fig. 3.4) can be derived from Van den Broek and Maarleveld (1963).

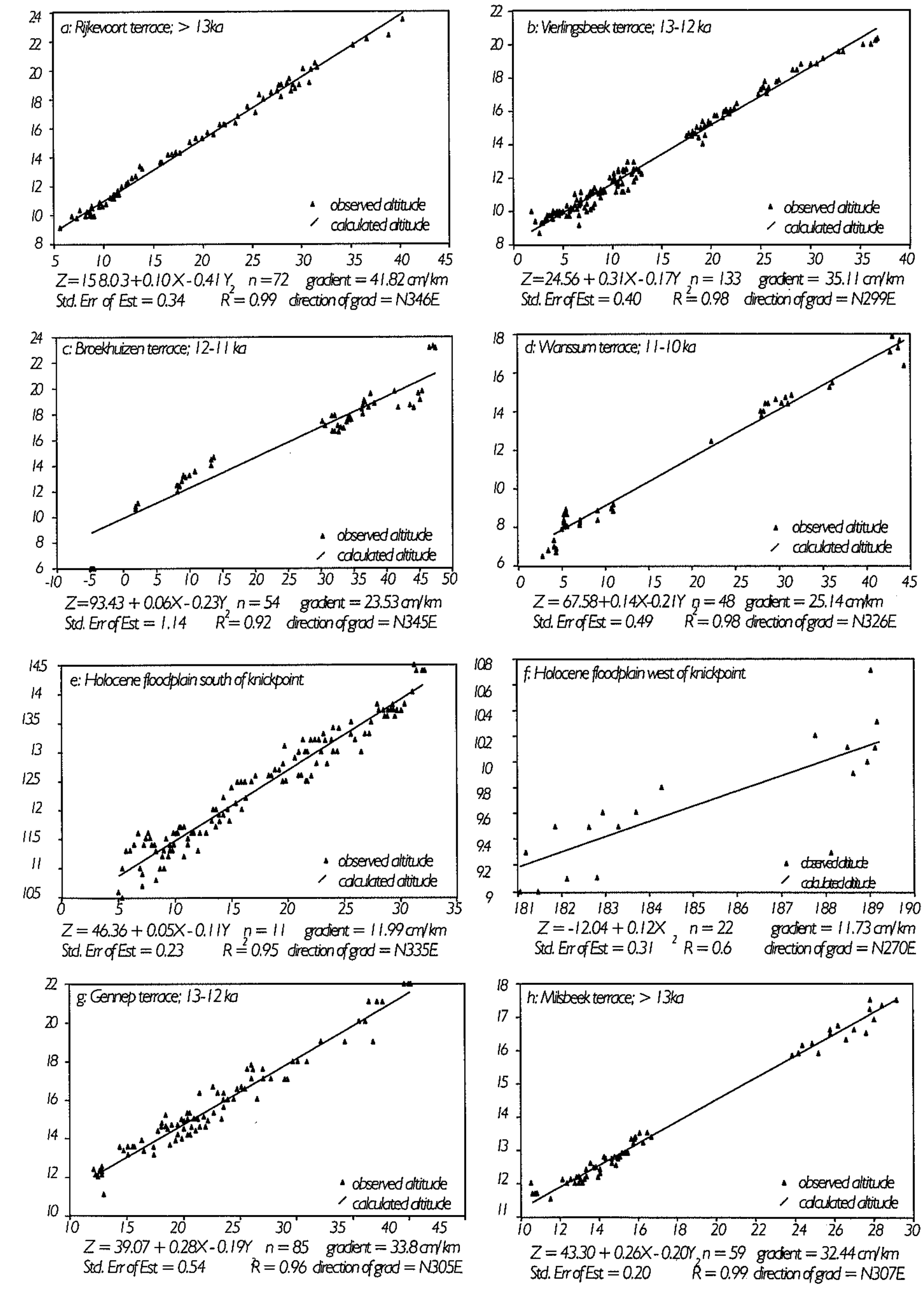


Figure 3.3 Multiple linear regression curves of Late Weichselian Maas and Niers terraces and Holocene Maas floodplain. Vertical axis is the observed altitudes in m above Dutch ordinal altitude (NAP), horizontal axis is horizontal distance upstream from arbitrary datum in km.

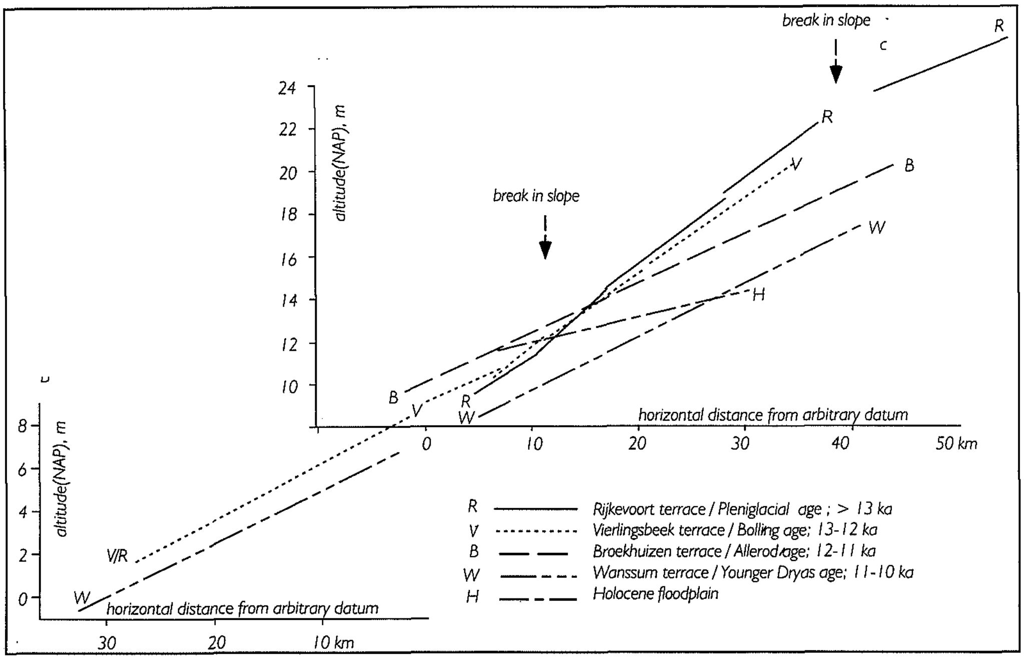


Figure 3.4 Compilation of longitudinal profile data for the Weichselian and Holocene terraces along the River Maas (data from this study, inset A); with data for the upper reach (C) of the Rijkevoort terrace from Van den Broek and Maarleveld 1963, and data from the lower reach (inset B) from Berendsen et al. 1995.

In order to see where exactly on the Vierlingsbeek- and Rijkevoort terraces the break in slope occurs, a subdivision of the data into four separate data sets is made. From each data set, consisting of 34 (Vierlingsbeek terrace) and 17 (Rijkevoort terrace) altitude observations, a separate mlr analysis has been made (Table 3.1). The location of these reaches is on Fig. 3.2. The accuracy of the analyses decreases and the R squared varies between 0.6 and 0.9, since the amount of data used in the analysis diminishes. The most downstream reach of the Vierlingsbeek terrace (reach 1, Table 3.1) has a slope of 27,0 cm/km, which matches the value observed downstream by Berendsen et al. (1995). The slope increases drastically to 39,6 cm/km from the first to the second data set (reach 2, Table 3.1), and is maintained in the upstream reaches (36,2 cm/km and 34,8 cm/km). The Rijkevoort terrace also shows an increased steep slope value in the second data set of 46 cm/km, which decreases upstream into 38 cm/km for the third and fourth data reaches (Table 3.1). It is interesting to see the change in direction of the terrace slope as well: the most upstream reaches show a south-north direction, whereas the more downstream reaches indicate a change in direction towards the north-west.

3.5 TECTONIC ACTIVITY INFERRED FROM GRADIENT LINES

One way to explain the break in slopes on the oldest terraces (Fig. 3.4) is to consider the confluence of the Niers. The Niers was comparable in size with the Maas, considering the palaeo-valley dimensions, until the Allerød, when it stopped serving as part of the Rhine river system (Huisink 1997). The extra input of sediment might explain the lower terrace slopes downstream of the confluence, but does not explain the break in slope more upstream (Fig. 3.4), where a gentler value slope is seen. The position of this break in slope of the Rijkevoort- and Vierlingsbeek terraces is located where the Maas crosses a fault which marks the boundary between the Peel Block and the Venlo Graben (Fig. 3.1).

reach 1 of Rijkevoort terrace	*Z= 96.25+0.15X-0.28y Rsquared = 0.79 slope = 31.59 cm/km Std. Error of Est. = 0.27 no. of observations = 18 direction of slope = N331E*
reach 2 of Rijkevoort terrace	*Z= 152.91+0.17x-0.42y Rsquared = 0.96 slope = 45.75 cm/km Std. Err. of Est. = 0.21 no of observations = 18 direction of slope = N338E*
reach 3 of *Rijkevoort terrace*	*Z= 138.42+0.11x-0.36y Rsquared = 0.96 slope = 37.56 cm/km Std. Err. of Est. = 0.28 no. of observations = 18 direction of slope = N343E*
reach 4 of Rijkevoort terrace	*Z= 162.53+0.02x-0.37y Rsquared = 0.97 slope = 37.45 . Std. Err. of Est. = 0.26 no. of observations = 18 direction of slope = N356E*
reach 1 of Vierlingsbeek terr.	*Z= -69.45+0.26x+0.08y Rsquared = 0.61 slope = 27.64 Std. Err. of Est. = 0.34 no. of observations = 34 direction of slope = N286E*
reach 2 of Vierlingsbeek terr.	*Z= -39.47+0.39x-0.06y Rsquared = 0.90 slope = 39.64 Std. Err. of Est. = 0.28 no. of observations = 33 direction of slope = N278E*
reach 3 of Vierlingsbeek terr.	*Z= 38.05+0.30x-0.20y Rsquared = 0.90 slope = 36.20 Std. Err. of Est. = 0.45 no. of observations = 33 direction of slope = N304E*
reach 4 of Vierlingsbeek terr.	*Z= 155.45+0.002x-0.35y Rsquared = 0.99 slope = 34.8 Std. Err. of Est. = 0.18 no. of observations = 33 direction of slope = N360E*

Table 3.1 Regression analyses for the Vierlingsbeek (13-12 ka) and Rijkevoort (< 13 ka) terraces, where each terrace is divided into four data sets, location of each reach is on Fig. 3.2.

The "steeper" slope segments and breaks in slope of the two oldest terraces are probably best explained by concerning tectonic movements in this area. The area is part of the Roer Valley rift system, which has been subjected to tectonic movements for a long time. The Roer Valley Graben for instance has been subjected to differential subsidence from the Oligocene onwards. Displacements along the Peel Boundary Fault were 0.8 mm/a during the Quaternary (Geluk et al. 1994). Displacements still occur, which is illustrated by the earthquake of April 13, 1992. The faults related to the Peel Block (Fig. 3.1) are locally seen in the landscape as scarps, for instance in the northern part of this area where 7 m difference in altitude marks the boundary between the Rijkevoort terrace in the Venlo Graben and the Peel Block (Huisink 1997).

Tectonic movements of either the rising Peel Block or the subsiding Venlo Graben may possibly have led to tilting of blocks in the Venlo Graben, causing small scarps more or less perpendicular to the main faults. This is in accordance with a lineament analysis by Van den Berg (1994 b), who showed that the main faults are orientated in a NW-SE direction but a second, important direction, namely NE-SW is also seen by lineaments occasionally with signs of horizontal movements (Fig. 3.1). Tilting is not unknown in this area as tilting of the Peel Block is demonstrated by Van den Berg (1994 b). The river possibly reacted by incising upstream of such a scarp and depositing sediment downstream which resulted in a relatively steeper gradient in the area near the scarp in comparison with gradients in up-and downstream, tectonically unaffected parts of the valley. The preservation of such scarps is poor in this energetic fluvial environment. The tectonic activity responsible for the local steepening of the Rijkevoort and Vierlingsbeek terraces was apparently active before the forming of the Broekhuizen terrace (Allerød), since the longitudinal profile of this terrace can be correlated downstream without breaks in slopes.

3.6 IMPACT OF TECTONIC MOVEMENTS VERSUS CLIMATIC CHANGES ON FLUVIAL STYLES

The slopes of the Broekhuizen- and Wanssum terraces (resp. 23.5 and 25.4 cm/km) are quite similar. The palaeo-river styles, however, are distinctly different: a highly sinuous meandering river in contrast with a braided river. This is probably the most convincing evidence that the Lateglacial Maas did not change its pattern due to tectonic influences. The braided river system of the late Pleniglacial is recognized in large parts of the Maas valley, both up- and downstream (Huisink 1997, Berendsen et al. 1995, Van den Berg and Schwan 1996). Changes in channel dimensions or sediment characteristics have not been observed, regardless of differences in slope values. This also shows the minor influence of longitudinal profiles on the fluvial pattern.

The channel morphology on the Vierlingsbeek terrace (Bølling) changes downstream of Beugen, between reaches two and three (Fig. 3.2). The location of this change, however, does not coincide with the break in slope of the longitudinal profiles, which was found between the first and second reach (Table 3.1). The channel patterns on these parts of the terrace are identical, regardless of the large difference in terrace slope (39.6 versus 27.0 cm/km). Locally steeper longitudinal terrace profiles, related to tectonic movements, did probably not account for the change in morphology therefore.

A more plausible explanation for the change in morphology on the Vierlingsbeek terrace was the effect of the Niers confluence, a tributary of the Rhine river system at that time. Large amounts of water and especially sediment were added to the Maas system which is shown in the heavy mineral composition of the sediment (Huisink 1997). Extra input of sediment, especially in a transitional river, might be the cause for the more braided-like pattern downstream of Beugen. So the influence of sediment yield on river morphology seems to be far larger than the influence of longitudinal terrace profiles on river morphology in the Lateglacial Maas valley. The effect of the locally steeper terrace profile resulted only in deeper incised channels on the Vierlingsbeek terrace. Detailed sections across channel scars on the steeper sloping part of this terrace show deeper incised channels (up to 6 m, Huisink 1997) as compared to channels on gentler sloping parts of the terrace (up to 3 m).

The importance of tectonic movements, inferred from longitudinal terrace profile analyses, was most probably not large on the river evolution of the Maas during the Lateglacial. Although the uplift of the hinterland of the Maas on a Quaternary time scale resulted in the tendency to erode, and the forming of terraces, it could not account for the development of four terraces in such a short period as the late part of the Pleniglacial and Lateglacial. The phases of incision and deposition are better explained by the distinct climatic changes during the Lateglacial which affected sediment and water supply significantly.

The change in pattern from a braided river in the Pleniglacial via a transitional system into finally one, highly sinuous meandering channel in the Allerød is explained by reduced sediment load and regularization of discharge. The warming of the Lateglacial resulted in an increasingly denser cover of vegetation, which in turn diminished sediment load. The combination of permafrost thaw and vegetation development increased soil water storage capacity, which, combined with decreasing importance of snow melt as major supplier of discharge, resulted in a more regular discharge regime. The Maas reacted on the reduced sediment yield and more regular discharge regime compared to the previous period by incision, whereby the Rijkevoort terrace was formed. Minor channels became abandoned, while larger ones incised and became more curved. This trend proceeded until finally one highly sinuous channel remained active in the Allerød, when a closed wooded vegetation was present. During this process the Vierlingsbeek terrace (transitional system) was thus formed.

At the onset of the cold Younger Dryas an incision occurred due to a rapid change in discharge and a delay in vegetation disappearance, whereby bank stability was maintained. Later on the vegetation changed and an opener landscape (yielding higher sediment load) together with irregular discharges resulted in a braided river pattern (Wanssum terrace). Another incision took place at the onset of the Holocene when vegetation changed fast into a closed forest in response to higher temperatures. Discharge became much more regular and a meandering river evolved which incised the former floodplain. The meandering river remained confined to a narrow floodplain in which the river shifted and deposited merely fine sediments.

3.7 CONCLUSIONS

The changes in steepness of terrace slopes did not influence the Lateglacial Maas river styles. For instance, the longitudinal terrace profiles of the highly sinuous meandering river of the Allerød and the braided river of the Younger Dryas are more or less identical, while the morphology is distinctly different. The four phases of incision and accumulation that occurred since the onset of the Lateglacial were most probably not triggered by tectonic activity, but can be better explained by climatically induced factors.

Local tectonic activity, expressed by changes in steepening of the longitudinal terrace profiles on the Rijkevoort and Vierlingsbeek terraces did, most likely, not lead to changes in channel morphology. Instead, changes in channel pattern observed on the Vierlingsbeek terrace were probably caused by an extra input of sediment by a tributary. River styles could be correlated through large parts of the valley, regardless of local steeper terrace slopes. The local tectonic activity did possibly lead, on the other hand, to locally deeper incised channels in areas with steeper terrace profiles. Faults also influenced the flow direction of the river, which is more or less perpendicular to the faults.

Longitudinal terrace profiles, calculated by using multiple regression analysis, prove to be useful in the morphological mapping of terraces. They can be used for correlation purposes by connecting isolated terrace fragments.

4 Lateglacial river sediment budgets in the Maas Valley; The Netherlands

by M. Huisink. Accepted for publication in Earth Surface Processes and Landforms

4.1 ABSTRACT

Three Weichselian Lateglacial (13-10 ka) terraces have been distinguished in the Maas valley which were formed when the Maas repeatedly incised in an increasingly narrower floodplain. The River Maas changed from a braided system (before circa 12.5 ka) via a transitional phase to a high-sinuosity meandering river (circa 12.5-11 ka), to a braided system (circa 11-10 ka) again and finally to a low-sinuosity meandering river (after 10 ka). These fluvial style changes involved phases of erosion and deposition. The amounts of eroded, deposited and reworked sediment during each Lateglacial period will be calculated in this paper. The sediment budgets allow to compare the transport capacity of the different river styles, which will help to explain the observed fluvial changes. Borehole information regarding the thickness of terrace sediments and lateral extensions of the Lateglacial terrace surfaces were combined in a three-dimensional approach, using a GIS. Multiple regression analyses were used in calculating altitudes of entire terrace surfaces from individual altitude measurements. It will be shown that the fluvial development of the Maas can not only be explained by climate-related external factors such as sediment-discharge ratios and discharge characteristics, but possibly also by intrinsic factors such as floodplain dimensions and the channel morphology of previous periods.

4.2 INTRODUCTION

This paper aims to explain the complex nature and causes of fluvial responses to climate change by calculating amounts of eroded, reworked and deposited sediment by different river types during Lateglacial warm and cold periods. The Maas changed from a braided river in the Late Pleniglacial via a transitional phase into a meandering river in the Allerød, it became braided in the Younger Dryas and meandering in the Holocene (Bohncke et al. 1993, Vandenberghe et al. 1994, Kasse et al. 1995, Huisink 1997). Incision and terrace formation took place at climatic transitions, from a warm to cold period (Allerød to Younger Dryas) and cold to warm periods (Pleniglacial to Bølling, Younger Dryas to Holocene), whereby the Maas became confined in an increasingly narrower floodplain. The study reach of the Maas valley (Fig. 4.1) comprises the most complete Lateglacial terrace sequence and is located just upstream of the intersection of Weichselian and Holocene deposits.

The Lateglacial (13-10 ka) was a period of large climatic changes in North-Western Europe (Bohncke and Wijmstra 1988, Van Geel et al. 1989, Walker et al. 1994, Paus 1995, Hoek 1997a, Isarin 1997). Fig. 4.2 shows mean annual and July temperatures for the Lateglacial and the division in the Bølling-Allerød interstadial and Younger Dryas stadial, which is commonly used in The Netherlands. The Pleniglacial represents the middle part of the last glacial of which the Late Pleniglacial (circa 27 - 13 ka) was the period of maximum cold. The impact of these climatic changes on lowland rivers in North-Western Europe is recognized in many river valleys (e.g. Kozarski 1983, Schirmer 1983, Rose 1995, Kalicki and Zernickaya 1995, Antoine 1997). The Maas valley in particular, being one of the largest rivers of The Netherlands, has been intensively studied by Pons and Schelling (1951), Van den Broek and Maarleveld (1963), Bohncke et al. (1993), Vandenberghe et al. (1994), Berendsen et al. (1995), Kasse et al. (1995), Van den Berg (1996) and Huisink (1997, 1998).

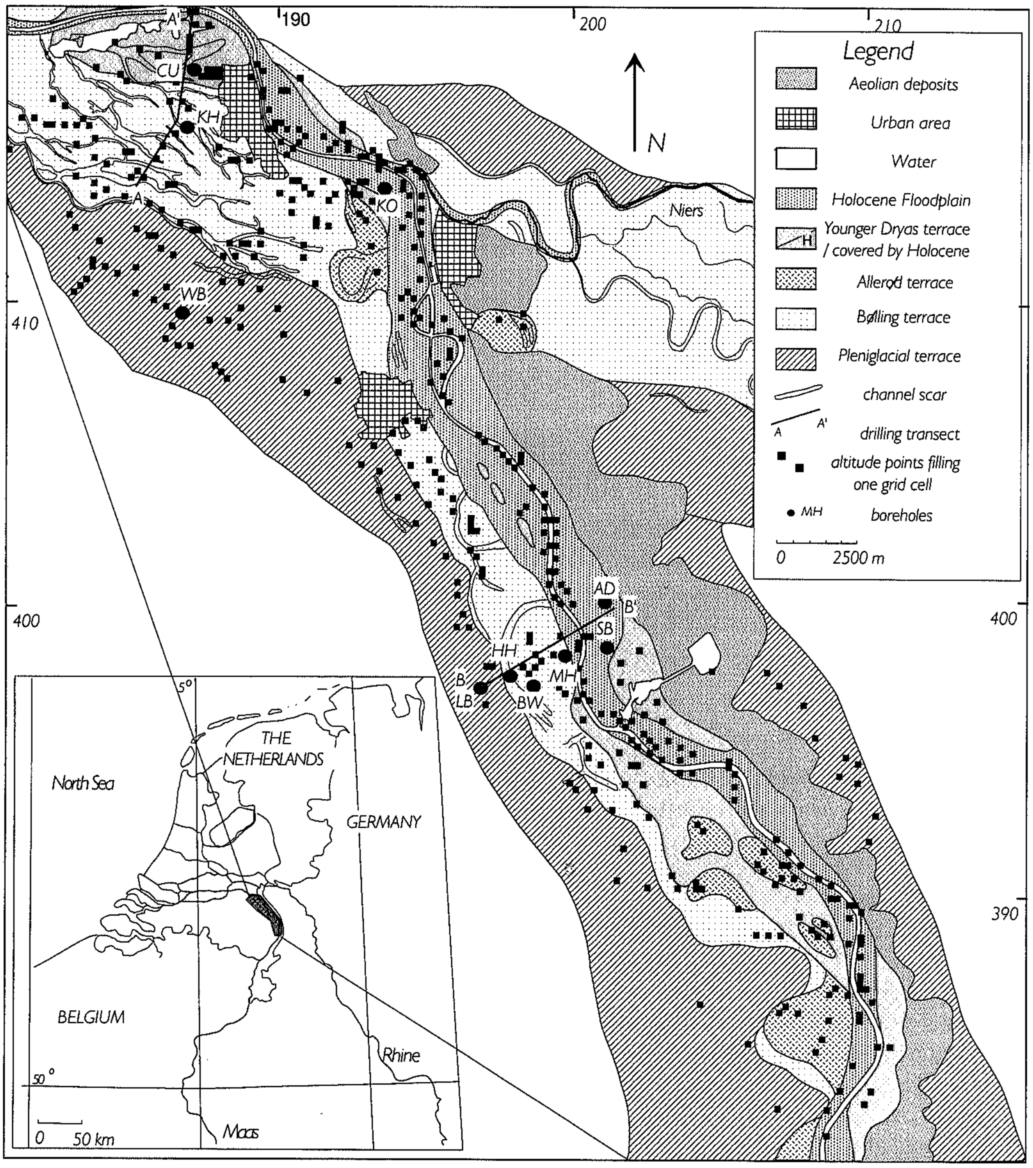


Figure 4.1 Location of study area and Lateglacial terraces.

The River Maas drains an area of 33.000 km^2, is rainfed and discharges range from 300 m^3/s in summer to 3000 m^3/s in winter.. Fluvial style changes of the Maas during the Lateglacial are most probably related to the distinct climatic changes in that period. The occurrence of tectonic activity did not account for the observed fluvial changes, although it influenced the formation and preservation of terraces in the Maas valley (Huisink 1998). The hinterland of the Maas, the Ardennes has been subjected to a gradual uplift from the Miocene onwards with periods of accelerated tectonic activity superimposed on a general continuous trend (Van den Berg 1996). The long term uplift of the Ardennes cannot explain the formation of three terraces in 3000 years however. Lateglacial tectonic activity in the Venlo Graben, in which the study reach is located, is shown by steepening of the longitudinal profiles of parts of the Pleniglacial and Bølling terraces, but it did not trigger fluvial changes (Huisink 1998). The braided (late Pleniglacial) and transitional (Bølling) river systems could be traced along large parts of the valley without changes in morphology or sedimentology, regardless of (tectonically induced) locally steeper slopes,. The meandering Allerød river and the braided Younger Dryas river had different river styles but similar floodplain gradients (respectively 23,5 cm/km and 25,1 cm/km) which shows that floodplain gradients were not determining the fluvial style. Sea level was not an important control since it was some 100 m lower in the Lateglacial (Jelgersma 1966) and the study area was at 400 to 500 km distance from the coastline at that time.

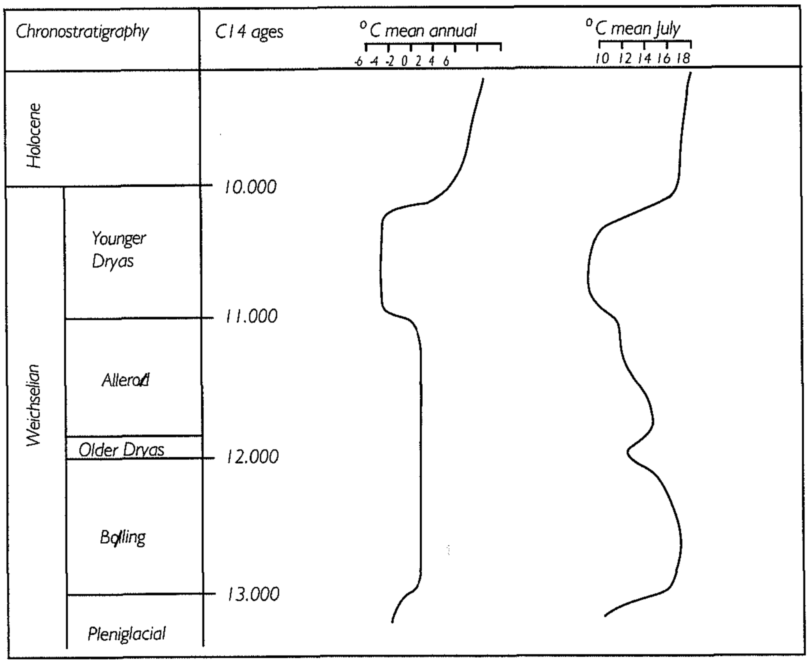
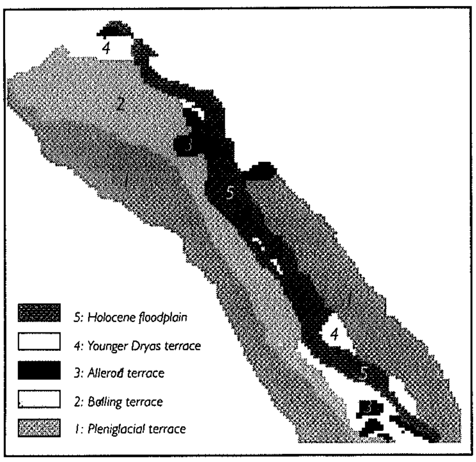


Figure 4.2 Chronostratigraphy and temperatures in the Weichselian Lateglacial. Mean July temperature after Walker et al. 1995; mean annual temperature after Kasse et al. 1995a.

Since the terrace extensions are known and mechanically drilled boreholes provide information about sediment thickness, it is possible to quantify the erosion, deposition and reworking of the Lateglacial River Maas. Quantitative fluvial studies often deal with short time-scales and site-specific river reaches where a lot of data are gathered (e.g. Lane and Richards 1997) or to modeling of large areas over long time periods where limited detailed data are available (e.g. Veldkamp and Van den Berg 1993). Quantification of erosional and depositional phases during the last glacial usually involve one-dimensional estimates of depths of incision or deposition from particular sites. By using a GIS, a three-dimensional approach is possible in a valley stretch. The GIS enables to make calculations and shows the geographical variation within the study area, as the fluvial processes of erosion and sedimentation change from up- to downstream parts in the valley. The distinction between the terraces and the sedimentology of the sediment facies have been described previously by Huisink (1997). New data from deeper boreholes are used to determine the thickness of sediment units.

4.3 THE LATEGLACIAL TERRACE STRATIGRAPHY

One late Pleniglacial (circa 27 - 13 ka) and three Lateglacial terrace surfaces have been recognized (Huisink 1997). Sediment characteristics and channel morphology on the terrace surfaces enabled an interpretation of changing river styles. The oldest sediments in this study are deposited by a braided river in the cold Pleniglacial. The river occupied a large floodplain, bounded by higher relief in the west (Peelhorst) and remnants of older Rhine terraces in the east. The braided river system developed during the warmer Bølling period into a transitional style between braiding and meandering and the floodplain width decreased. Minor channels were abandoned, while larger ones started to incise, formed levees and deposited overbank sediments. The incision trend of major channels and abandonment of smaller ones continued until finally a single meandering channel remained active in the Allerød (12-11 ka, Fig. 4.1). This channel was confined to a rather narrow floodplain in which it moved laterally. At the onset of the cold Younger Dryas (11 ka, Fig. 4.1) the river incised significantly and changed abruptly to a braided river system. The river filled the incision only partially. New incision took place at the start of the Holocene and the Maas became meandering again. During the Holocene the river deposited mostly fine-grained sediments.



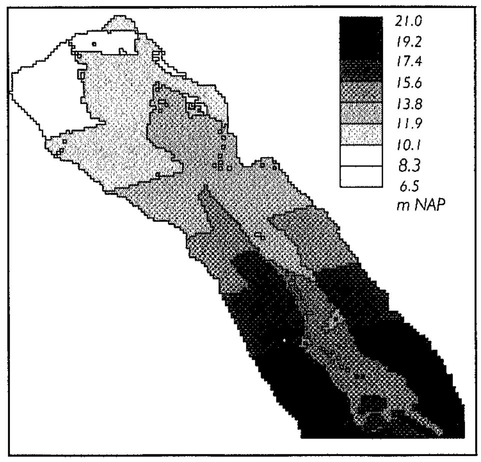


Figure 4.3 a The Lateglacial terrace map (based on figure 4.1) ; 4.3 b The Maas valley altitude map.

4.4 CREATION OF TERRACE AND ALTITUDE MAPS USING A GIS

Fig. 4.1 shows the extent of the terraces in the study area which is used as a base for the creation of GIS maps in PCRaster (Wesseling et al. 1996). This part of the Maas valley was divided in grid cells of 200 by 200 m, which allows an acceptable representation of the Maas valley. For each grid cell x- and y-coordinates, terrace code and altitude are given.

Altitudinal points (minimum of 54 and maximum of 133) on each terrace surface were obtained from 1:10.000 altitude maps. These points (Fig. 4.1) were carefully selected from the highest parts of the terrace surfaces, mostly the top of point bar sediments or levees (Huisink 1998) and used in multiple linear regression analyses to obtain the terrace altitudes in the whole valley reach. The output of the multiple linear regression analyses is an equation for each terrace, that represents the best fit of a surface through the observed altitude points. These equations enable the altitude of each grid cell in the Maas valley to be calculated. The accuracy of the multiple linear regression analyses proved to be high as the R squared varied from 0.92 to 0.98 (Huisink 1998). GIS maps of the present River Maas valley are represented in Fig. 4.3, showing the terrace code in each grid cell (Fig. 4.3a) and the calculated altitude (Fig. 4.3b).

4.5 MAXIMUM AND MINIMUM TERRACE DIMENSIONS

PCRaster is used to calculate palaeo-floodplain surfaces in the late Pleniglacial and Lateglacial before erosion or incision by adding subsequently incised terraces. The surface of former floodplains is difficult to establish when remnants of younger terrace surfaces are scattered or only located on one side of the river. In the latter case the range between maximum and minimum extensions of palaeo-terrace surfaces is calculated.

4.5.1 Pleniglacial floodplain dimensions

As the Pleniglacial terrace surface is found on both sides of the River Maas (Fig. 4.1), the Pleniglacial river must have occupied all the intermediate area as well (Fig. 4.4a).

4.5.2 Bølling floodplain dimensions

The Bølling terrace is found only on the west side of the River Maas. The maximum floodplain area is the combination of the Bølling terrace surface with all younger terrace surfaces (Fig. 4.4b). A calculation of the minimum Bølling floodplain area is the sum of the terrace remnants that are actually found in the valley (Fig. 4.4c), which is not the most probable assumption since incision of this terrace took place in the Allerød (Huisink 1997, Kasse et al. 1995a). The average of the minimum and maximum calculated floodplain sizes is in this case more realistic.

4.5.3 Allerød floodplain dimensions

The reconstruction of the Allerød palaeo-floodplain dimension is difficult since terrace fragments are scarce. Most of the terrace sediments were eroded during the Younger Dryas and Holocene. A maximum floodplain area is the sum of the Allerød and Younger Dryas terrace surfaces and the Holocene floodplain (Fig. 4d). A minimum estimation of the Allerød floodplain area, based on present-day terrace remnants only, has not been made because the Allerød terrace fragments do not form a continuous floodplain (Fig. 1). A better estimate of the minimum floodplain size of the Allerød meandering river can be obtained by comparison with the actual, also meandering river floodplain, since the size of the Allerød meander scars is comparable with meanders in the Holocene floodplain (Fig. 1).

4.5.4 Younger Dryas floodplain dimensions

Most of the Younger Dryas terrace sediments have been eroded in the Holocene but Younger Dryas terrace remnants are found on both sides of the Holocene floodplain (Fig. 4.1). The maximum terrace surface is therefore obtained by adding the Younger Dryas and Holocene floodplain dimensions

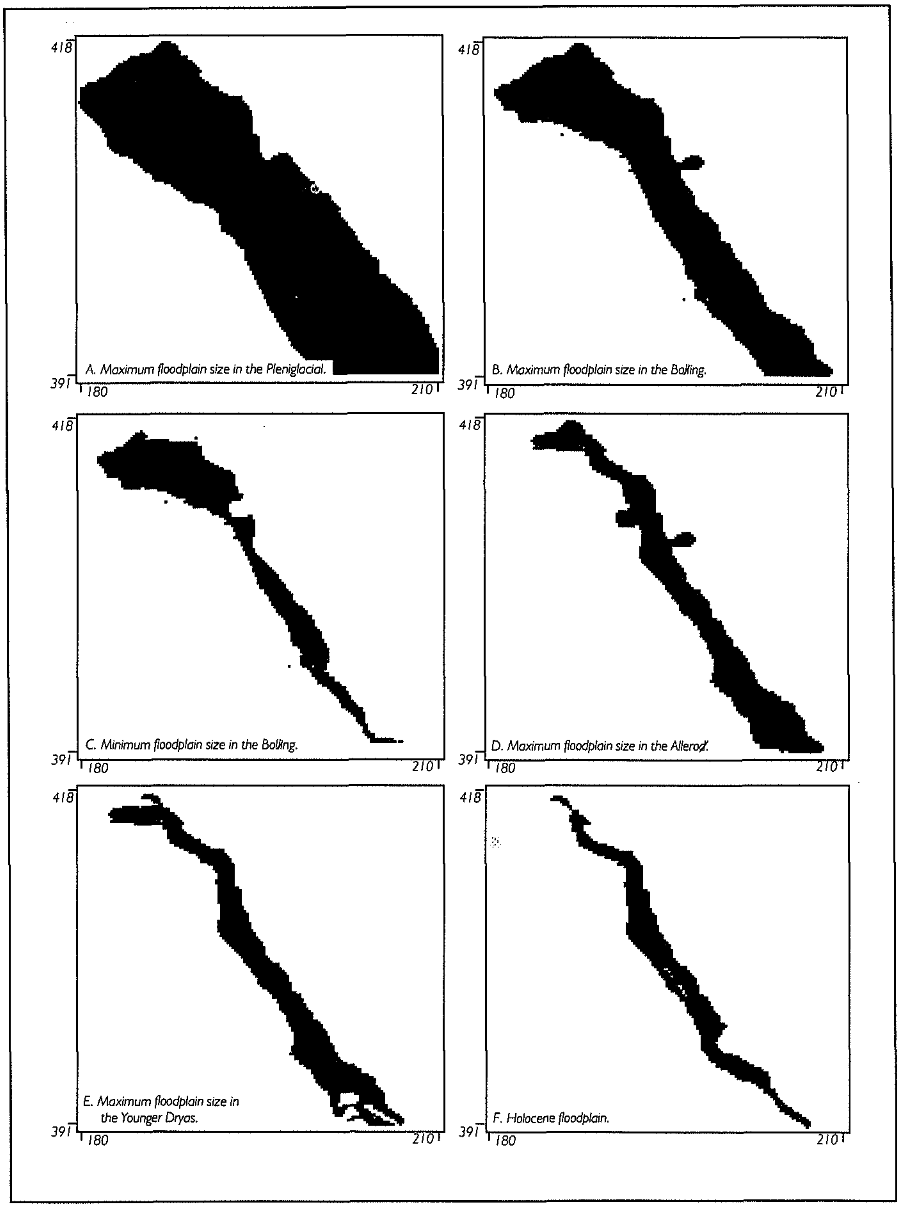


Figure 4.4 Lateglacial floodplain dimensions.

(Fig. 4.4e). The minimum floodplain size cannot be obtained because the Younger Dryas terrace is found as fragments only and the terrace surface is covered by Holocene deposits in the northern part of this study area. The maximum extension of the Younger Dryas floodplain is the most accurate reconstruction possible and is therefore used in this study.

4.6 THICKNESS OF SEDIMENT BODIES

The late Pleniglacial and Lateglacial terrace sediments consist mostly of coarse sand and gravel and are described in detail by Huisink (1997). Table 4.1 shows a summary of sediment characteristics which have been used to differentiate Lateglacial terrace sediments from older deposits and to estimate terrace sediment thickness. Most boreholes in this area have been made by hand and did not penetrate the basal gravels of the terrace sediments (Huisink 1997), so mechanical boreholes were used that penetrated deeper and provided undisturbed samples. Six boreholes were located in the southern part of the study area, on transect B-B' (Fig. 4.1, 4.5) and four in the northern part, partially on transect A-A' (Fig. 4.1, 4.5). Lithology, sedimentological structures and grain sizes are provided in Fig. 4.6a and b.

terrace	*sediment characteristics*	*length of fining-up sequences*
Holocene floodplain	*mostly fine-grained, sometimes clayey sand, silt and organic material. lowermost sediments are coarse*	
Younger Dryas	*coarse-grained, poorly sorted, gravelly sand. abrupt lithological changes, both laterally and horizontally.*	*- 2 m*
Allerød	*channel fill: peat and gyttja fine-grained, rather well sorted, silty sands at the top of a fining-up sequence, changing down - wards into coarse, poorly sorted gravelly sands*	*3 m*
Bølling	*top facies: fine to medium gravelly sands, gradual litholo-gical changes*	*2 - 3 m*
	lower facies: poorly sorted, medium to coarse sand and gravel, abrupt lithological changes	*not distinct*
Pleniglacial	*poorly sorted, medium to coarse gravelly sand with grainsize changing rapidly in vertical and horizontal direction*	*mostly not distinct 0,5 - 1 m*

Table 4.1 Sediment characteristics of the Maas terraces.

4.6.1 The thickness of the late Pleniglacial terrace sediments

The base of the terrace sediments varies from 5.9 m (boring LB, Fig. 4.6a) to 7.5 m (boring WB, Fig. 4.6a) below the terrace surface. In borehole LB a distinct change in lithology from coarse, gravelly, poorly sorted sand to predominantly fine-grained, well sorted sand with silt and clay layers marks the boundary between late Pleniglacial terrace sediments and older sediments. This is confirmed by the heavy mineral composition of samples lb 32 and lb 17 (Fig. 4.6a, table 4.2) as sample lb 17 matches the mineral composition of the late Pleniglacial terrace sediments (Huisink 1997) with a high amount of stable and opaque minerals and a relatively low amount of hornblende and volcanic minerals, while sample lb 32 shows a higher amount of hornblende and volcanic minerals and a lower amount of stable minerals. In borehole WB the lithology changes in the hiatus between 7.4 and 8.4 m, at circa 8 m, where coarse-grained sands and gravels without calcium carbonate are replaced by clayey, fine-grained calcareous sands and gravels. The terrace sediments are, however, covered with 0.5 m of aeolian sands, so that the thickness of the fluvial terrace deposits is 7.5 m in borehole WB.

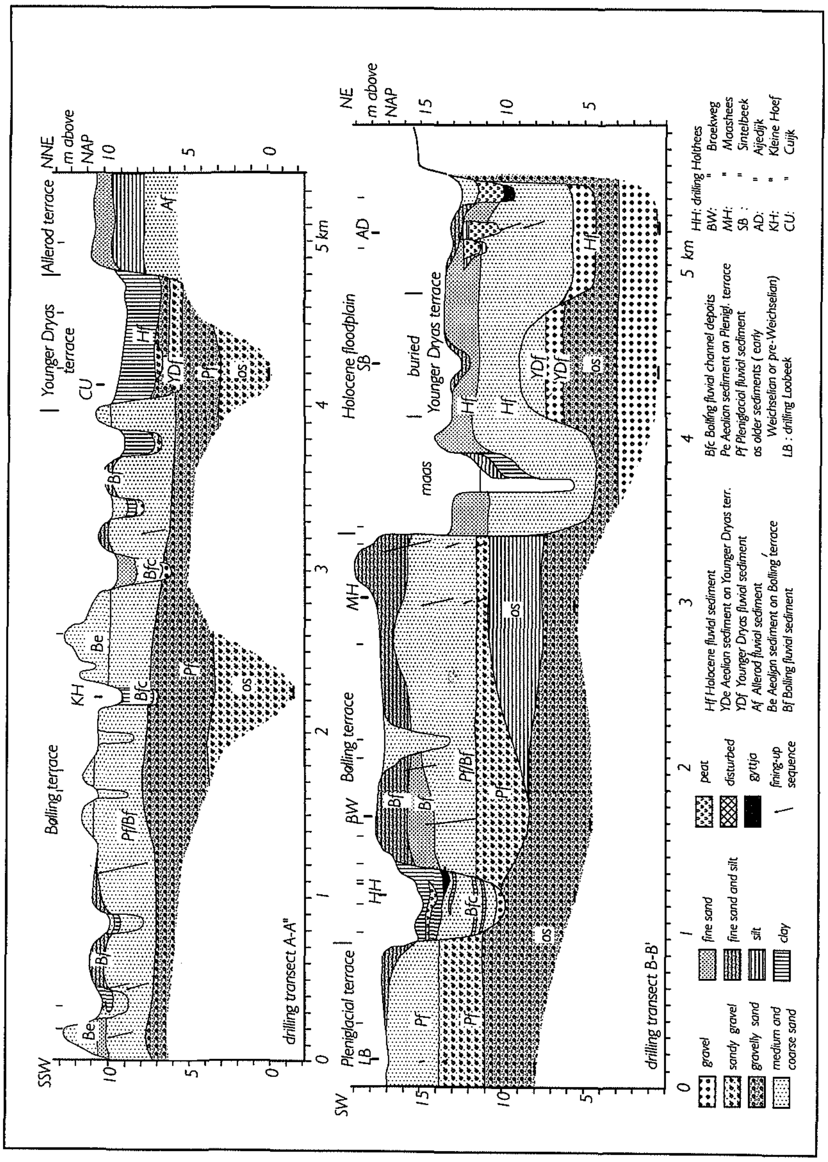


Figure 4.5 Borehole transects A-A' and B-B' (for location see figure 4.1).

4.6.2 The thickness of the Bølling terrace sediments

The Bølling terrace sediments were deposited on top of the Late Pleniglacial river sediments. The change in fluvial style from the Pleniglacial to the Lateglacial did not involve the complete erosion of the braided river sediments but only reworking of the top sediments in the active floodplain of the transitional river system. The coarse sediments of the late Pleniglacial braided river and the lower facies of the transitional Bølling system cannot be distinguished (table 1, Huisink 1997). The upper facies of the Bølling terrace (table 4.1) is therefore used to distinguish Bølling sediments from older deposits which is at 5.9 m in borehole MH and at 6 m in borehole BW.

Boreholes HH and KH are located in channel scars and used to determine the depth of incision in the channels on this terrace surface by channel lag deposits. Near borehole HH (Fig. 4.1), several small gullies merge in one large, curved channel (Vandenberghe et al. 1994), which is typical for the transitional river system. The minimum depth of this channel is 5 m and is formed by a gravelly lag deposit. The thickness of Bølling sediments in borehole KH is most likely at 3,3 m where a gravelly lag deposit is found. In order to estimate the total amount of erosion between the late Pleniglacial and Bølling, the depths of the channel fills in these boreholes have to be added to the altitude differences between the top of the channel fills and the terrace surface. This results in 7.1 m incision for borehole HH and 4,3 m for KH.

4.6.3 The thickness of the Allerød terrace sediments

The maximum depth of the meander scar south of borehole KO (Fig. 4.1) is 5 m from the base of the channel fill to the top of the older terrace surface in which the channel has incised and the minimum thickness of the Allerød sediments is around 3.5 m (Kasse et al. 1995a). Further south a maximum of 7m was found in another meander pointbar sequence (Kasse et al. 1995a), which is not used in these calculations since its position is outside the study area.

4.6.4 The thickness of the Younger Dryas terrace sediments

The Younger Dryas terrace sediments are found under Holocene sediments in the northern part of the area (boreholes CU and SB). The Younger Dryas sediments in borehole CU (Fig. 4.6b) are found from 2 to 4.9 m below the surface, and are characterized by distinct short fining-up sequences and usually medium-grained sand (< 600 mu). They contrast with the coarser underlying sand and gravel (> 600 mu) without distinct fining-up sequences. The top 2 m of this borehole consist of fine-grained Holocene overbank deposits. In borehole SB the top 3.6 m consist of fine-grained, silty Holocene floodplain deposits. From 3.6 to 6.3 m Younger Dryas sediments occur as mostly coarse gravelly sand, separated from the underlying, also coarse sediments by a gravel lag deposit. Two heavy mineral samples from the Younger Dryas sediment at 4.6 m (SB 20) and the underlying sediments at 7.5 m (SB 34) show subtle differences. The amount of opaque minerals in SB 20 (42%) points to a Younger Dryas sediment (Huisink 1997), while SB 34 contains somewhat high amounts of alterite, chloritoid and stable minerals and a lower amount of opaque minerals. The thickness of the Younger Dryas sediments is 2.87 m in boring CU and 2.63 m in boring SB.

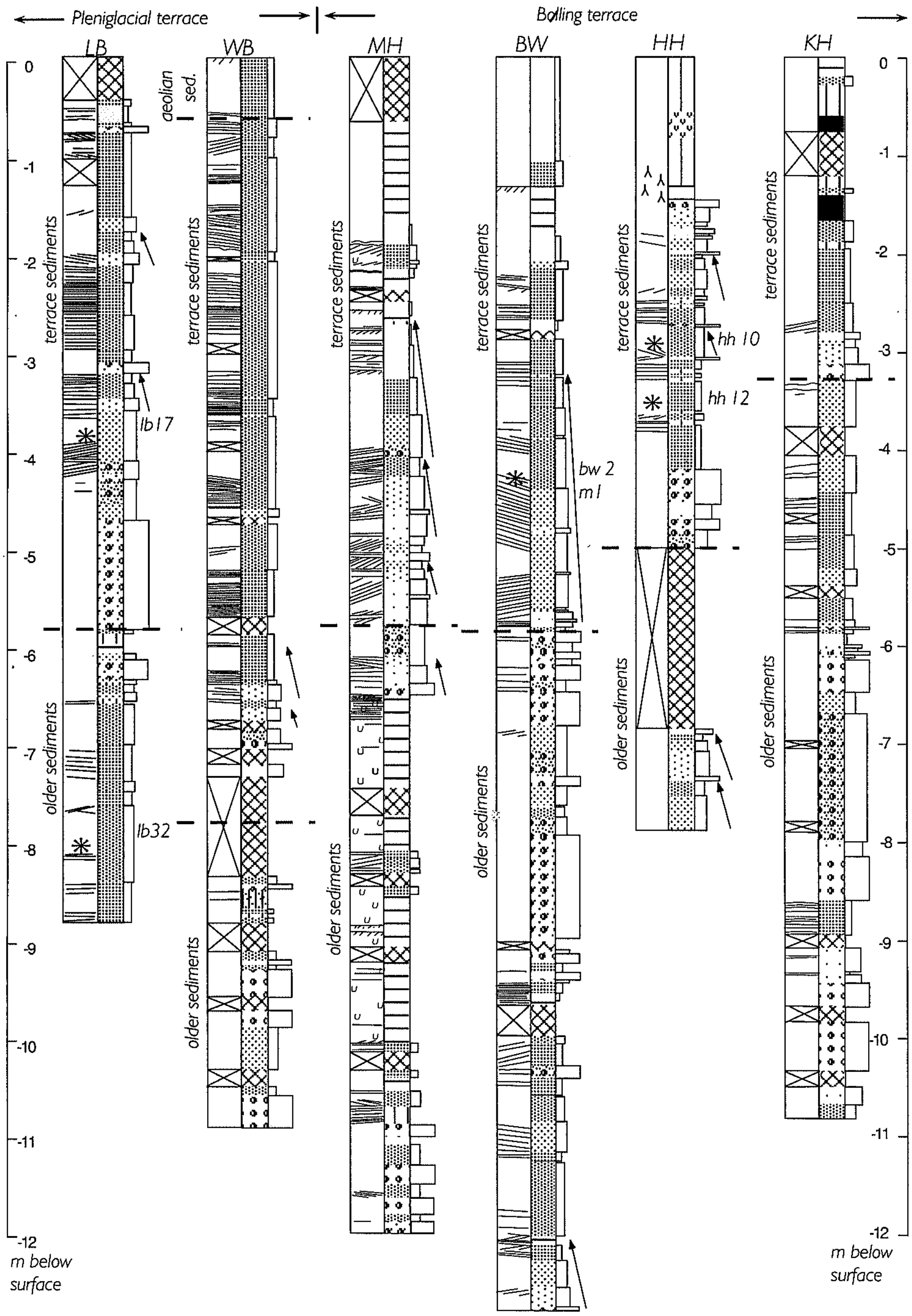


Figure 4.6a Borehole logs of boreholes LB, WB, MH, BW, HH and KH (for location see figure 4.1).

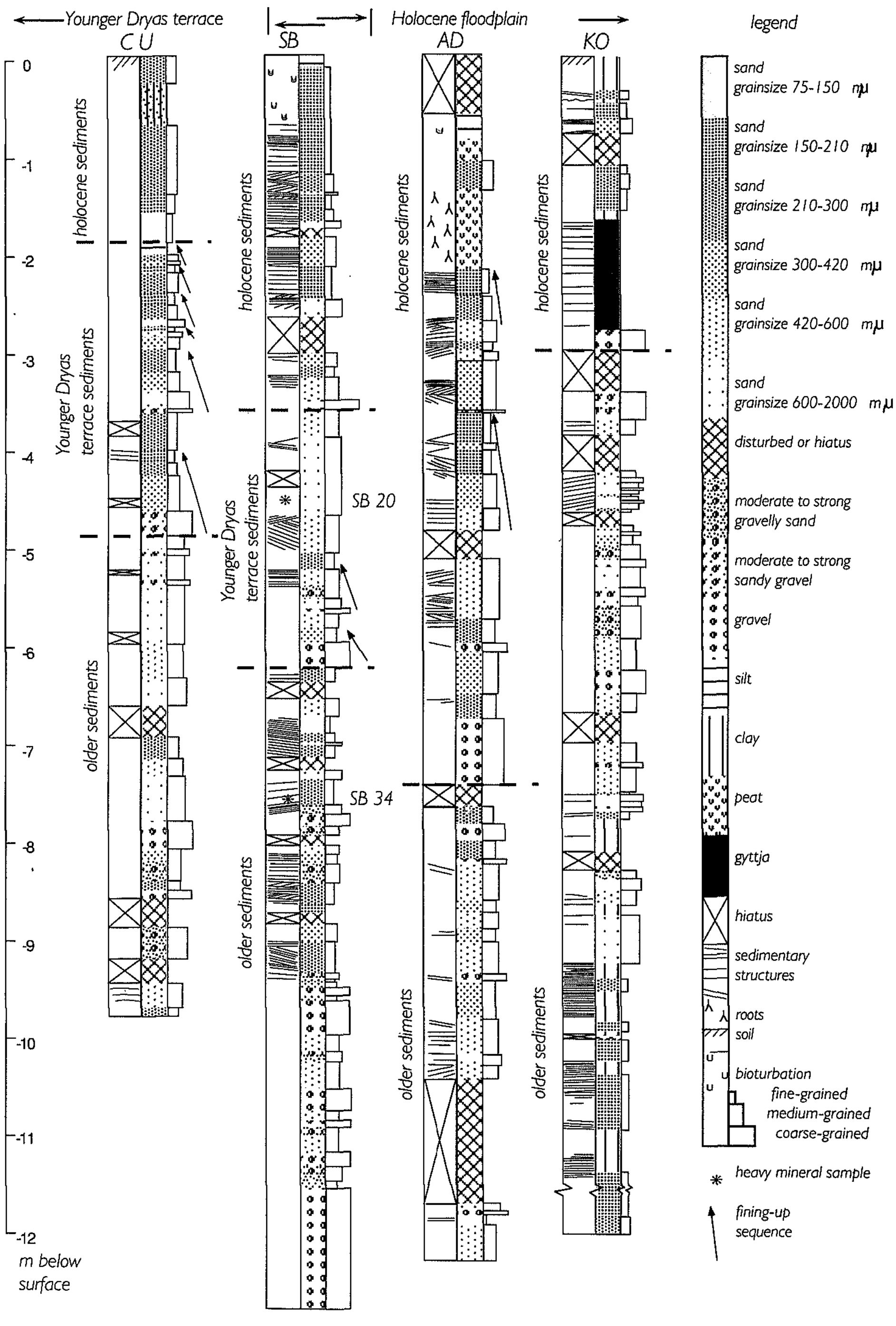


Figure 4.6 b Borehole logs of boreholes CU, SB, AD and KO (for location see figure 4.1).

4.6.5 The thickness of the Holocene floodplain deposits

Three boreholes were made in the Holocene floodplain: SB, AD and KO (Fig. 4.1), in which the depths of the Holocene sediments could be recognized by their mostly fine-grained nature or by their heavy mineral content. Hand borings were also used to estimate the thickness of the Holocene floodplain sediments because of the, mostly, fine-grained sediments which enabled a deep penetration. The maximum depth of the Holocene sediments is found in boring AD (Fig. 4.6b), where at 7.4 m a gravel layer forms the coarse base of the Holocene deposits. An overall fining-up trend is seen, and most of the sediments are fine-grained, silty or humic. The maximum depth of about 7 m is observed in hand boreholes as well (Huisink 1997). Two other mechanical borings, SB and KO (Fig. 4.6b), show shallower depths. A gravel layer at 3.6 m in boring SB forms the boundary between the upper fine-grained, Holocene deposits and somewhat coarser, gravelly Younger Dryas sediments which contain short fining-up sequences and a Younger Dryas heavy mineral composition (sample SB 20 Fig. 4.6b). The thickness of the Holocene deposits in KO (Fig. 4.6b) is at least 3 m as fine-grained humic deposits like gyttja abruptly overly a gravel layer which probably marks Holocene sedimentation after an erosional phase.

	lb17	*lb32*	*bw2m1*	*hh10*	*hh12*	*sb20*	*sb34*
garnet	*13.2*	*8,5*	*5,0*	*30,3*	*27,0*	*29,7*	*16,3*
epidote	*8,8*	*9,0*	*25,7*	*12,4*	*14,0*	*19,3*	*15,3*
alterite+saussurite	*6,9*	*7,0*	*8,4*	*4,5*	*3,0*	*5,0*	*15,8*
hornblende	*5,4*	*10,0*	*2,5*	*2,5*	*3,0*	*1,0*	*0,5*
chloritoid	*0*	*1,0*	*1,5*	*0*	*1,0*	*0,5*	*2,5*
volcanic minerals	*2,9*	*9,5*	*2,0*	*1,5*	*3,5*	*2,0*	*0,5*
stable minerals	*23,0*	*14,0*	*17,3*	*16,9*	*22,5*	*12,4*	*19,8*
unstable minerals	*2,5*	*1,0*	*5,0*	*2,5*	*3,0*	*1,0*	*0*
topaz	*0*	*0*	*1,0*	*0*	*0*	*0*	*0,5*
staurolite	*9,3*	*10,5*	*10,9*	*7,0*	*4,5*	*10,9*	*10,4*
metamorphic minerals	*8,8*	*11,5*	*5,9*	*7,0*	*7,0*	*4,0*	*2,0*
tourmaline	*19,1*	*18,0*	*14,9*	*15,4*	*11,5*	*14,4*	*16,3*
opaque minerals	*34*	*32*	*52*	*34*	*34*	*42*	*34*
in percentages (%)							

Table 4.2 Heavy mineral content of samples from boreholes (for location see figures 4.6 and 4.1).

4.7 RESULTS: LATEGLACIAL REWORKING OF SEDIMENT

The combination of terrace sediments thickness and terrace dimensions is used to estimate the amounts of reworked sediment in each Lateglacial period. All calculations are based on the sediment that is stored nowadays in the Maas valley. The results are net erosional and depositional budgets and are used to compare Lateglacial fluvial changes. The amount of sediment transported through this part of the Maas valley during the Lateglacial is not estimated here. The maximum amount of reworked sediment is the maximum sediment depth times the maximum palaeo-floodplain size and the minimum amount is the minimum depth times the minimum floodplain size (Fig. 4.7).

The maximum amount of reworking in the Holocene ($4.10^8 m^3$) is comparable with that in the Allerød ($4{,}5.10^8 m^3$), but the reworking in the Younger Dryas ($8{,}3.10^8 m^3$) is twice as large and reworking in the Bølling ($1{,}2.10^9 m^3$) is even three times as large. The minimum amounts of reworked sediment in the Holocene and Allerød are equal ($1{,}6.10^8 m^3$), while the amounts in the Bølling and Younger Dryas periods are considerably higher (respectively $3{,}4.10^8 m^3$ and $4{,}7.10^8 m^3$). In this scenario, the reworking in the Younger Dryas must have been larger than in the Bølling.

The average thickness is probably the most reliable value to use in the calculations, as the thickness of sediment bodies varies. The average thickness is combined with the average palaeo-floodplain size for the Bølling, the minimum floodplain size for the Allerød and the maximum floodplain size for the Younger Dryas as these represent most likely the palaeo-flood plains best. When the most realistic estimates of reworked sediment during the Lateglacial are compared (Fig. 4.7), the smallest amounts of reworking occurred in the Allerød ($2{,}1.10^8 m^3$), 1,4 times as much in the Holocene ($2{,}9.10^8 m^3$), 3 times as much in the Younger Dryas ($6{,}3.10^8 m^3$) and 3,2 times as much in the Bølling ($6{,}8.10^8 m^3$).

Comparable amounts of reworked sediment are found for the Holocene and Allerød periods, while larger amounts are found in the Bølling and Younger Dryas periods. This is understandable since the river type during the Allerød and Holocene was meandering, while it was braided in the Younger Dryas and changing from braided to meandering during the Bølling. The amounts of reworked sediment in the Bølling and Younger Dryas are roughly the same, in spite of the differences in depths of incision between the Younger Dryas (up to 10,5 m) and the Bølling (6 -7 m). The amount of reworked sediment in the Bølling and Younger Dryas periods is three to four times as much as in the Allerød which is explained by the more dynamic rivers (respectively transitional between braided and meandering and braided) during these periods, in contrast with a meandering river in the Allerød. Although the amounts of reworked sediment in the Holocene and Allerød are comparable, the dynamics of the two rivers were not comparable at all, since the Allerød lasted for 1000 years and the Holocene for 10.000 years. The dynamics of the Holocene river is therefore some 10 times lower.

4.8 NET EROSION AND DEPOSITION IN THE LATEGLACIAL

For each terrace surface (minimum, maximum and most likely) an elevation map was made. Net erosion or deposition between the terraces were estimated by subtracting elevation maps of subsequent terraces. Fig. 4.7 shows the amounts of net erosion or deposition that were calculated by using the most likely terrace surface and the average sediment thickness. The minimum and maximum terrace surface areas were also used in the calculations to indicate the range in which the erosion or deposition occurred. Each grid cell had a higher or lower altitude after the subtraction. The overall altitude difference was estimated by adding up all grid cells of one terrace. This altitude difference was then divided by the total palaeo-floodplain area to obtain the average sedimentation or incision per square meter.

The change from the braided late Pleniglacial river into a transitional one between braided and meandering in the Bølling resulted in a net sedimentation. When the maximum and minimum possible Bølling palaeo-floodplain sizes are subtracted from the Pleniglacial floodplain the amount of sedimentation is $1{,}2\ m/m^2$ for the maximum size and $1{,}1\ m/m^2$ for the minimum size. The average floodplain size, which was considered the most likely floodplain size, results in a similar sedimentation.

The change from the Bølling to the Allerød is characterized by a change into one, meandering channel which eroded slightly. Four scenarios can be considered, namely the maximum or minimum Bølling surface being eroded by the maximum or minimum Allerød surfaces. All four calculations show erosion, ranging from 0,3 m/m^2 (min. Bølling surface minus both the maximum and minimum Allerød surfaces) to 1,6 m/m^2 (max. Bølling surface minus both the maximum and minimum Allerød surfaces). The average of these amounts, around 1 m/m^2 (Fig. 4.7) is probably the most realistic amount of erosion, since the average of the maximum and minimum Bølling floodplain sizes is the most reliable palaeo-floodplain size. Within the study area a difference is seen between the up- and downstream parts. For instance when the maximum Bølling surface was eroded by the maximum Allerød surface 3,7 m sediment per grid cell was eroded in the most upstream part, whereas 0.9 m sediment per grid cell was deposited in the most downstream part. This demonstrates that fluvial processes are related to the position of the valley stretch examined. Rose (1995) demonstrated the complex fluvial responses of the River Gipping where aggradational and incisional phases changed from the steep uplands to the low relief stretches.

The transition from the Allerød to the Younger Dryas period is marked by a deep incision on a floodplain scale (Huisink 1997). The incision was filled partially with sediment during the Younger Dryas, but the net result is erosion, ranging from 2,6 (maximum Allerød surface eroded by the maximum Younger Dryas surface) to 2,9 m/m^2 (minimum Allerød surface eroded by the minimum Younger Dryas surface). The most likely scenario, where the minimum Allerød surface is eroded by the maximum Younger Dryas floodplain cannot be calculated with accuracy but the range from 2,6 to 2,9 m/m^2 erosion is quite satisfactory to indicate the possible amounts of eroded sediment. Erosion occurred in the whole study reach.

The change from the Lateglacial to the Holocene is marked by incision at first and sedimentation afterwards (Huisink 1997), which filled the larger part of the Holocene floodplain. The most likely scenario where the Holocene floodplain is subtracted from maximum Younger Dryas surface shows an average sedimentation of 1,5 m/m^2. This is the amount of sedimentation on top of the Younger Dryas sediments. The incision at the start of the Holocene, however, was sufficient to erode most of the Younger Dryas sediments. The sedimentation during the Holocene was therefore much larger, at places up to 6 - 7 m (Huisink 1997). So the net sedimentation calculated in this study is an underestimation.

Sedimentation rates, the average amount of sedimentation per year during each Lateglacial period, are not estimated as these are not reliable. Organic material is scarce in the Lateglacial Maas sediments which makes it difficult to define the duration time for each erosional and depositional phase. Another complicating factor is that the sediment stored in the Maas valley is the net result of each period. What happened within this period is hard to pinpoint in time. The sediment could have been aggraded during the whole period, but also during the last 100 years, which influences sedimentation rates greatly.

4.9 DISCUSSION

The Maas changed gradually from a braided river in the Late Pleniglacial via a transitional phase into a meandering river at the onset of the Lateglacial. The transitional river was thought to be active from around 12,7 ka until 11,8 ka B.P. (Kasse et al. 1995a, Huisink 1997). However, according to Hoek (1997b) the dates around 12, 7 ka by Teunissen and De Man (1981) and Teunissen (1990) are some 500 years too old. A new date on macro remains from organic detritus in the upper sandy facies of the braided river sediments in the Bosscherheide pit revealed an age of 12,390 ± 100 yr. B.P. (GrA-9431). This means that the Maas was still braided in the early Bølling, until circa 12,4 ka B.P., although probably less dynamic than during the Late Pleniglacial. Afterwards the larger channels incised and became curved forming the multichannel transitional phase as described by Vandenberghe et al. (1994). Although the larger channels incised up to 7 m locally, a net aggradation occurred on a floodplain scale (see Fig. 4.7). The first net incision on a regional scale occurred by the single-channel meandering river that was active from the late Bølling or early Allerød onwards.

The incision and change in fluvial style during the Bølling can be explained by changes in discharge - sediment ratios. The warming of the Lateglacial resulted in a vegetation development from an open tundra towards shrub tundra in the early Bølling, towards increasing amounts of birch copses in the Bølling sensu stricto and eventually into an open birch forest in the early Allerød (Hoek 1997a). This will have reduced slope erosion rates greatly (Kirkby 1980) which resulted in prograding incision of the channels. Discharges in the Bølling were highly irregular, related to snow melt, because winter temperatures remained low in the Bølling period. As the soils were still frozen when snow began to melt in springtime, the soil water storage was low, and discharges will have had a peaked character. The result was a highly dynamic fluvial environment as is shown by the large amount of reworked sediment in this period (see Fig. 4.7) which indicates a high eroding and transporting capacity. The vegetation development towards the end of the Bølling period resulted in a significantly reduced sediment load which led to incision on a regional scale. The vegetation development will have increased evapotranspiration rates as well which led to diminished discharges. The river became confined into a narrow floodplain in the Allerød when bank stability was high due to the dense vegetation cover.

This slow response to warming was quite different from the abrupt fluvial response to the Younger Dryas cooling. The Maas incised significantly on a regional scale most probably during the early Younger Dryas (Kasse et al. 1995a, Huisink 1997) and changed its pattern towards braided without a transitional phase. Aggradation is thought to be active at least from 10.500 yr BP onwards as aeolian dune accumulation occurred (Bohncke et al. 1993) and the dune sand came from the palaeo-floodplain (Huisink 1997). The amounts of sediment that were reworked by the Younger Dryas braided river were comparable with those of the Bølling transitional river. Erosion occurred, however, in a different manner. Bed erosion took place in a confined, narrow floodplain and sedimentation afterwards was not enough to compensate for the erosion, while erosion in the Bølling period occurred in a wide palaeo-floodplain, by shallow channels and erosion was balanced by sedimentation.

Discharges at the Allerød - Younger Dryas transition changed fast due to cooling and wetting of the climate (Bohncke et al. 1987, 1993) becoming larger and more irregular, while the vegetation, consisting of an open birch forest with pines (Hoek 1997a) forming a good protection against bank and soil erosion (Huisink 1997) remained intact at first. The increased discharge - sediment ratio resulted in erosion. The vegetation changed during the Younger Dryas as the forest vegetation became more open from 10.950 BP onwards and the vegetation became less dense in the second half of the Younger Dryas (Hoek 1997a). This reduced bank and soil stability and increased sediment yields resulting in aggradation. The amounts of reworked sediment and the sediment itself in the Younger Dryas and Bølling periods are similar which probably indicates that the stream power was comparable. The Younger Dryas river was however confined into a narrow floodplain and bank stability was higher due to the more dense vegetation cover. The excessive stream power was therefore spent on bed erosion, rather than lateral erosion and the configuration of the Younger Dryas floodplain was thus determined largely by the confinement of the river into a narrow floodplain in the previous Allerød period.

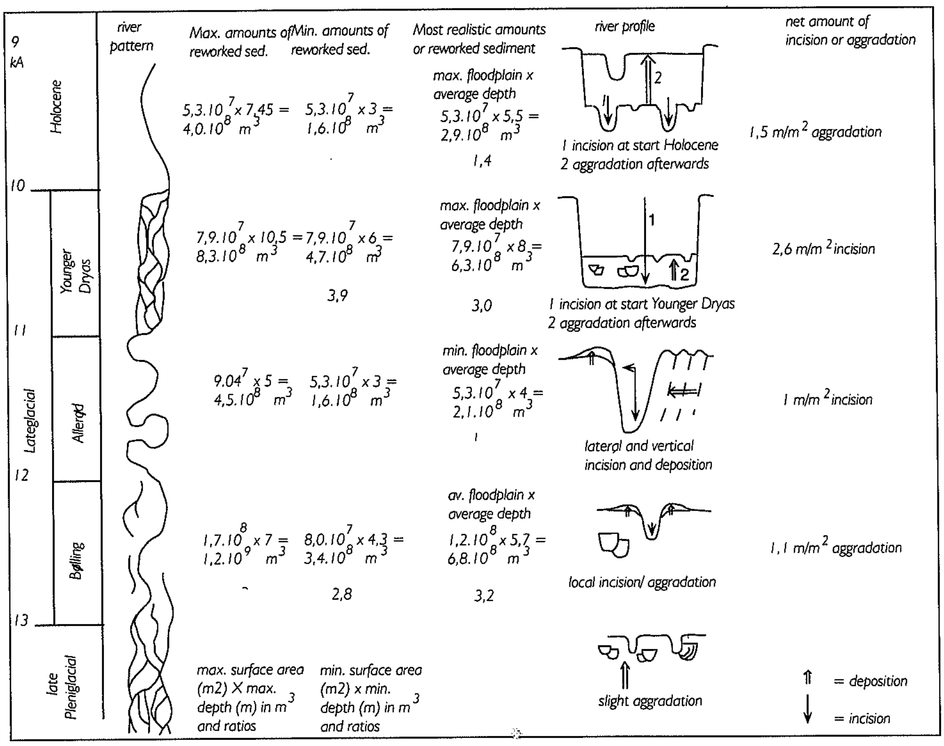


Figure 4.7 Synthesis of river pattern, reworked sediment and net erosion or deposition during the Lateglacial.

The Maas incised at the onset of the Holocene warming and changed into a meandering river but, in contrast with the Bølling period, without a transitional phase. Incision took place at the very beginning of the Holocene, in the Preboreal (Kasse et al. 1995a) and probably dominated until accumulation became dominant in the Atlantic period (Berendsen et al. 1995). In contrast to the onset of the Lateglacial a dense birch wood vegetation quickly established (Hoek 1997a), thereby reducing the sediment load. The rise of winter temperatures reduced the importance of snow melt and discharges became more evenly distributed throughout the year. The fast changes in water and sediment supply accelerated the fluvial style change.

Although a closed vegetation cover was present in the Allerød and Holocene, and in both periods the summer temperatures were high, a much higher amount of sediment was reworked during the Allerød period, namely 5 to 10 times more. This can be explained by the colder winter temperatures in the Allerød. The cold winters resulted in a frozen soil in spring and thus in a low soil water retention capacity when discharges were high. The Allerød discharges were dependant mostly on snow melt and therefore strongly seasonal. This resulted in a high stream power and a high dynamic meandering river that eroded, transported and deposited a large amount of sediment. Although the fluvial style of the Holocene and Allerød periods was similar, being both meandering, the Allerød river was much more dynamic and more comparable in stream power with the other Lateglacial river styles.

The Maas reacted to Lateglacial warming and cooling events by incision and changes in fluvial style. This is conform the model of Vandenberghe (1993, 1995) who proposed a non-linear response of a river to climate-related changes in water and sediment supply. These changes were significant enough to force the river to respond and to cross geomorphic thresholds (Schumm 1977). Kozarski (1991) suggested already a strong forcing of the vegetation cover on river development as changes in vegetation cover and river styles coincided in time. Rose (1995) explained Lateglacial channel changes of the River Gipping by changes in run-off and sediment supply (related to vegetation cover and soil development) but stressed the complexity of fluvial processes as these vary from up- to downstream parts of the catchment. Although only a small part of the floodplain is studied in this paper changes between up-and downstream parts could already be observed. This is most evident in the calculated erosion of the Allerød period, when incision occurred in the upstream part, but deposition in the most downstream part.

The high amounts of reworked sediment during the Lateglacial show that these rivers were highly energetic, and had large stream powers which can be explained by the specific discharge characteristics. The low winter temperatures resulted in a frozen soil at the time of snow melt in spring, so soil water storage capacity remained low and discharges were highly irregular and peaked, resulting in a temporarily large stream power. The depth and rate of incision was strongly correlated with the vegetation cover development. The calculations in this paper showed that net incision on a regional scale occurred when a woody vegetation was present. Apart from external factors influencing the development of the Maas, an internal factor like configuration of the palaeo-floodplain is thought to be important as well. The confinement of the Maas into a small floodplain in the Allerød resulted in the deep incision in the Younger Dryas period.

4.10 CONCLUSIONS

The Lateglacial warming resulted in diminished reworking of sediment from the late Pleniglacial to the Allerød and a change in fluvial style from braided via a transitional phase to meandering. During the Bølling net sedimentation took place on a floodplain scale, regardless of incision of major channels on a local scale. Net incision occurred during the Allerød, by a meandering river, when sediment supply was relatively low due to a closed and woody vegetation cover. The cooling and wetting at the start of the Younger Dryas resulted in a change in fluvial style from meandering to braiding. The amount of reworking became four times as much as in the previous period and a net incision occurred. The change in pattern at the onset of the Holocene into a meandering river resulted in a much less dynamic river compared to the rivers in the Lateglacial. The amount of reworking in the Holocene is 5 to 10 times less compared to the meandering river in the Allerød and about 20 times less compared to either the Younger Dryas or Bølling. These differences in river dynamics are related to the much more regular discharge regime and the complete cover of vegetation in the Holocene. This study shows that the Lateglacial configuration of the floodplain was strongly controlled by climate forcing (external factor), but also by the architecture of the floodplain in a previous period (internal factor).

5 The heavy mineral and gravel composition of Lateglacial Maas sediments

5.1 INTRODUCTION

Heavy mineral- and gravel analyses were performed on Lateglacial fluvial deposits as part of a morphological and sedimentological study of the River Maas (Chapter 2). In order to understand the sediment characteristics that are found in this part of the Maas valley in the Venlo Graben (see Fig. 5.1), the pre-Lateglacial sedimentation history has to be considered as well.

From the early Pleistocene on the Rivers Maas and Rhine deposited sediments in the southern part of The Netherlands (Zagwijn 1967). During the Tiglian the Maas was a tributary of the Rhine which deposited Rhine dominated sediments (the Tegelen Formation) in the Venlo Graben. During the Middle Pleistocene the Rhine changed its course towards the east outside the study area (Van den Toorn 1967). The River Maas incised at the beginning of the Holsteinian and deposited sediments afterwards in the Venlo Graben (Veghel B Formation). Tectonic activity at the beginning of the Saalian Pleniglacial (Van den Toorn 1967, Stiboka 1975) led to a subsidence of the Central- and Venlo Graben (Fig. 3.1, chapter 3). The River Maas migrated from the now relatively high Peelhorst into the lower lying Venlo Graben in the east. Incision into older, Rhine derived sediments in the Peelhorst area led to a mixed Rhine-Maas assemblage in the Venlo Graben. Fresh derived Rhine sediments were deposited in the Venlo Graben during the Saalian Pleniglacial when the Rhine was blocked by ice-masses in the north and was forced to the west into the Maas valley. Sediments from this period (Veghel-C Formation) consist therefore of a mixture of Rhine and Maas sediments. During the early Weichselian the Maas incised and most of the Veghel-C sediments in the Venlo Graben were eroded. The Rhine had a more eastern course, but one Rhine branch was situated in the present-day Niers valley (Fig. 5.1) and supplied Rhine derived sediments in the Maas valley until the Weichselian Lateglacial. During the Weichselian the Maas filled the Venlo Graben with sediments of the Kreftenheye B and C Formation (Van den Toorn 1967). Reworking and mixture with older, Rhine influenced sediments was important in the Venlo Graben and the sediments therefore have both a Rhine and Maas mineralogy.

The complex fluvial history of the Venlo Graben in this area (Fig. 5.1), resulted in a fill of both Maas and Rhine derived sediments. During the Lateglacial the Maas changed gradually from a braided river into a meandering one (Huisink 1997, Kasse et al. 1995a, Vandenberghe et al. 1994). The river became confined to one channel which reworked much less sediment than the braided river in the previous period (chapters 2 and 4). As a result it might be expected that a relative enrichment of fresh supplied Maas sediments would occur in the deposits during the Lateglacial, as reworking of older Maas-Rhine sediments diminished. The deep incision on a floodplain scale during the Younger Dryas (Huisink 1997) whereby older sediment was eroded and reworked is possibly visible in a change in heavy mineral or gravel composition. The main aim of this chapter is to establish if the Lateglacial fluvial style changes of the River Maas and the associated phases of incision and deposition are reflected in the gravel or heavy mineral composition. In this respect, the confluence of a Rhine branch (in the Niers valley) into the Maas should not be neglected.

5.2 METHODS

5.2.1 Gravel countings

From 24 samples the gravels in the 3-5 mm fraction have been analyzed. The first samples have been counted twice, to check the reliability of the countings. All sample sites are located in the confluence area of the Maas and Niers (Rhine) but a distinction is made between Maas and Niers terraces (table 1, Appendix A Table 5.1). The average gravel content for each terrace can be found in Table 5.2.

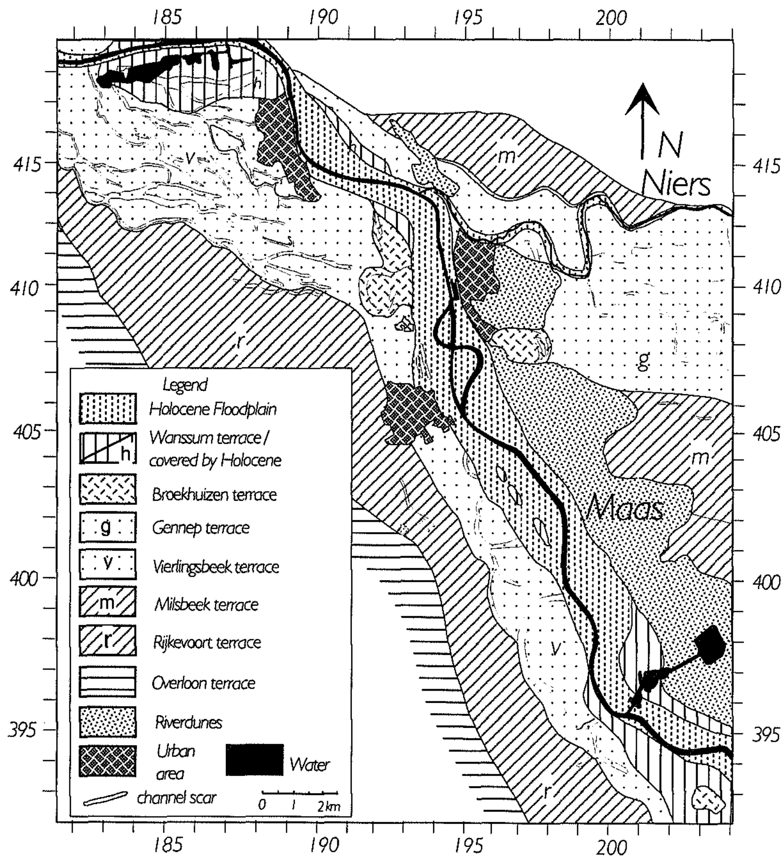


Figure 5.1 Lateglacial terraces in the Maas and Niers valleys.

5.2.2 Heavy minerals

From the fluvial and aeolian sediments in the Maas and Niers valleys 124 samples have been examined for the heavy mineral composition (Appendix A, Table 5.2). Although a variety of grain sizes occurs in the sediments, most samples are fine- to medium grained sands (150-300 mu). In the laboratory all samples were sieved for the fraction 50-420 mu. This is in accordance with the standardized grain size class according to Boenigk (1983) and is widely applied in other heavy mineral studies (e.g. Van Huissteden 1990, Krook 1993). This enables a good correlation with other heavy mineral studies. A standardized grain size is important since for instance Burger (1970) showed that the percentages of tourmaline, metamorphic minerals, saussurite, alterite and volcanic minerals increase with larger grain size and consequently, the amounts of zircon, rutile and garnet diminish with larger grain size.

Chronostratigraphy			*Maas terraces*	*Niers terraces*
Holocene		*Subatlantic*		
		Subboreal		
		Atlantic		
		Boreal	*Holocene floodplain*	*Holocene floodplain*
		Preboreal		
Weichselian	*Late Glacial*	*Younger Dryas*	*Wanssum terrace*	
		Allerød	*Broekhuizen terrace*	
		Bølling	*Vierlingsbeek terrace*	*Gennep terrace*
	late Pleniglacial		*Rijkevoort terrace*	*Milsbeek terrace*
pre-Weichselian/ Saalian			*Overloon terrace*	

Table 5.1 The chronostratigraphic position of the Maas and Niers terraces.

The medium to divide the heavy from light minerals was bromoform, with a specific gravity of 2.89 g/cm^3. From each sample 200 transparent minerals were counted. A distinction is made between samples located in the Maas valley up- and downstream of the Niers (Rhine) confluence and upstream of the confluence in the Niers valley (Table 5.4 and 5). A distinction is also made between fluvial and aeolian sediments. First results from heavy mineral analyses were published before (Huisink 1997).

The heavy minerals are arranged into the following groups: garnet, epidote, alterite, hornblende, chloritoid, volcanic, stable rest, unstable rest, metamorphic, topaz/staurolite and tourmaline. Opaque minerals are expressed in percentages of the total mineral content (opaque and transparent). The garnet group contains garnet, red garnet and turbid garnet. The epidote group contains epidote, klinozoisite, zoisite and piedmontiet. The distinction between epidote and alterite is subjective since epidote alters to alterite and all kinds of transitional grains can be found. The epidote and alterite percentages should be treated with caution therefore when a comparison is made with other petrographical studies. The alterite group comprises alterite and saussurite. The chloritoid group contains both chloritoid and turbid chloritoid. The volcanic minerals that have been recognized in this study were: basaltic hornblende, augite, enstatite, hyperstheen, titanite, anthophyllite and olivine. The stable rest group contains zircon, rutile, anatase, brookite, spinel, dumortherite, monazite and xenotime. The unstable rest group is composed of tremolite, apatite and diopside. Topaz and staurolite are taken into one group merely to limit the amount of groups, the amount of topaz being very limited (see Appendix A, Table 5.2). The metamorphic minerals are kyanite, andalousite, sillimanite, fibrolite and korund. Tourmaline has been counted as a separate group. This division is based on the mineral groups of the former State Geological Survey with some small modifications and may be used to distinguish Maas from Rhine sediments.

5.3 GRAVEL COMPOSITION OF LATE WEICHSELIAN MAAS AND NIERS TERRACES

5.3.1 Results

The sample from the Overloon terrace (pre-Weichselian) differs from the other samples in a high percentage of milky quartz, feldspar and quartz with feldspar, while the percentage of transparent quartz and greygreen sandstone is low (Table 5.2). According to Zonneveld (1947) this sediment belongs to the "Veghel zone" of Saalian age. The Maas terraces differ from the Overloon terrace in a lower amount of milky quartz and a higher amount of transparent quartz. The average amount of quartz (64%) in the Maas sediments is lower than from the Overloon terrace sediment (68%) but higher than the average amount of quartz in the Niers sediments (55%). This difference in total quartz content is merely the result of a lower amount of milky quartz in the Niers terrace sediments (Table 5.2). The Maas terraces have a lower amount of greygreen sandstone (average of 26%) when compared with the Niers terraces (average of 33%) and, less distinct, a lower amount of lydite. The greygreen sandstone gravels consist mostly of well rounded Paleozoic grey to greygreen sandstones which are indicative for Rhine sediment (Verbraeck 1984). High amounts of milky quartz seem to be typical for Maas sediments.

The difference in gravel composition between the Niers- and Maas terraces is largest in the youngest sediments while the oldest terrace deposits (Milsbeek- and Rijkevoort terraces) are quite similar (Table 5.2). The similarity in gravel content of both Pleniglacial terraces is probably the result of a thorough mixing of sediments by the highly dynamic braided rivers Maas and Niers (Rhine) in the confluence area (Huisink 1997). Apparently the mixing of Maas and Niers (Rhine) sediments declined during the Bølling period since the Vierlingsbeek and Gennep terraces show less similarity. This might be related to the change in fluvial style during this time. In the Bølling the larger channels started to incise as the rivers changed gradually from braiding to meandering, while smaller channels were abandoned. During this process mixing of Maas and Niers (Rhine) sediment in the confluence area decreased because the rivers generated separate flood plains and the gravel composition of the newly formed flood plains became increasingly characterized by either a Maas supply or a Niers (Rhine) supply. The gravel composition in the Holocene Niers floodplain differs significantly from the gravel composition in the Holocene Maas floodplain downstream of the confluence. This suggests that the present influence of the Niers sediment on the Maas gravel composition downstream of the confluence is small.

Within the Maas terraces an ambiguous trend is observed from old to young, namely a decrease in the amount of transparent quartz and an increase in milky quartz. This might be related to diminished reworking of older (mixed Rhine- and Maas) sediment and an increase of freshly supplied Maas sediments during the Lateglacial. The Wanssum terrace sediments deviate from this trend in the amount of milky quartz which is explained by the rather deep incision during the Younger Dryas which provided older, mixed Rhine- Maas sediments.

	Mix	Maas	Maas	Maas	Maas	Niers	Niers	Niers	aeolian
terrace or aeolian deposit	Over-loon	Rijke-voort	Vierlings-beek	Wanssum	Flood-plain	Mils-beek	Gennep	Flood-plain	river dune
N	1	1	4	2	3	6	5	1	1
milky quartz	55.8	39.9	46.4	44	47.8	40.4	37.8	37.7	42
red transparent quartz	1.2	0.5	2.9	2.6	1.9	2.3	2.4	1	1.3
transparent quartz	10.8	23.2	18.4	15.4	14	16.5	16.5	9.2	24.7
total quartz	67.8	63.5	67.7	62	63.8	59.2	56.7	47.9	68
flint	0.8	0.5	0.9	0.4	0.3	0.6	0.4	0	0
porphyry	0.3	0.5	0.7	1.1	1	1	0.6	1.3	0.7
feldspar	1.2	2.1	0.3	0.7	0.4	0.3	0.9	0.7	0.3
quartz+feldspar	2.8	0.5	1.1	0	1	0.5	1.3	0.7	0.7
crystalline	0	0	0.7	0	0.1	0.1	0.1	0.3	0
lydite	0.3	0	0.2	0	0.2	0.4	0.6	0.3	0
diverse	5.2	4.3	6.7	9.8	6	8.3	6.5	7.2	5.7
greygreen sandstone	21.5	28.7	21.7	25.8	27.2	29.7	29	41.3	24.7
limestone	0	0	0.1	0.2	0	0	0	0	0
total sandstone	26.7	33	28.4	35.6	33.2	38	38.4	48.5	30.3

Table 5.2. Average gravel compositions of the Maas and Niers terraces in percentages.

5.3.2 *Comparison with up- and downstream valley reaches*

A comparison with previous studies is difficult since different size classes have been counted, the distinction between quartzite and sandstone is subjective and quartz contents change downstream. Another complicating factor is the non-uniform division in gravel categories. Paulissen (1973) for instance divided his gravel in four groups: quartz, quartzite, sandstone and a rest group, while Maarleveld (1956) used eight groups: quartz, crystalline, flint, ringenkiezel, pyritic quartzite, oolite, lydite/radiolite and a rest group.

Paulissen (1973) analyzed samples in the more upstream part of the Maas valley, so these samples should have a pure Maas composition compared to the samples in the confluence area of Maas and Niers. In order to enable a comparison with Paulissen, my results were regrouped following Paulissen (Table 5.3). The Mechelen aan de Maas terrace can be correlated in age with the Rijkevoort terrace and the Geistingen terrace with the Vierlingsbeek terrace. The correlation between the two studies is poor. The amount of quartz and sandstone is much lower upstream, while the amount of quartzite is much higher. The different quartz contents might be explained by the different gravel size (8-16 mm) counted by Paulissen (1973). Maarleveld (1956) showed that a downstream decrease in gravel size is related to an increase in the total amount of quartz and a decrease in the amount of flint. This explains partly the relatively high percentages of quartz in the Rijkevoort and Vierlingsbeek terrace gravels. The difference in quartzite and sandstone content is probably due to the subjective division of the two gravel types.

The use of the percentage of quartz to indicate either a Maas or Rhine sediment source in 3 - 5 mm sized gravels is dubious. Maarleveld (1956) stated that the differences between the Maas and Rhine Saalian sediments in the 3-5 mm fraction are too small to differentiate on gravel composition between Maas or Rhine samples. Only in the coarser grained samples the Maas sediment can be distinguished by their higher flint content. Zonneveld (1947) stated that Maas sediments contain less quartz than Rhine sediments, which contradicts with the results from this study in table 5.2, where the Rhine derived Niers sediments contain lower amounts of quartz than the Maas samples. The grainsize of the analyzed samples are not mentioned by Zonneveld which made a correlation difficult. Hoek and Schorn (1990) analyzed Lateglacial gravels downstream in the "Land van Maas en Waal" and found a varying quartz content of 50 - 70 %, which is comparable with the results of this study. They concluded that the gravel composition is dominated by Rhine derived sediments. However, 50 to 70 % quartz can be found in both Rhine and Maas deposits (Maarleveld 1956). It is probably easier to distinguish Maas from Rhine sediments on the percentage of greygreen sandstone, since this seems to be significantly higher in Rhine derived sediments (Table 5.2).

	Peelhorst	Mechelen aan de Maas terrace	Rijkevoort terrace	Geistingen terrace	Vierlingsbeek terrace	Wanssum terrace	Holocene floodplain
quartz	67.8	30.5	63.5	27.7	67.7	62	63.8
quartzite	2	60.1	2.6	61.2	2	4.5	2
sandstone	26.7	5.4	33	6.7	28.4	35.6	33.2
rest	3.5	3.9	0.9	4.5	5.3	2.8	3.2
Gravel size	3-5 mm	8-16 mm	3-5 mm	8-16 mm	3-5 mm	3-5 mm	3-5 mm

Table 5.3. Comparison of the average gravel composition of Maas terraces in percentages from this study with the gravel composition of the "Mechelen aan de Maas" - and "Geistingen" terraces from Paulissen (1973).

5.4 HEAVY MINERAL COMPOSITION OF LATE WEICHSELIAN MAAS AND NIERS TERRACES

Typical Maas minerals in young Pleistocene deposits are zircon (stable minerals), tourmaline, turbid chloritoid and metamorphic minerals, while typical Rhine minerals are volcanic minerals like augite, hypersthene and titanite (Zonneveld 1947). Since the valley fill is a mixture of both Rhine- and Maas sediments it is expected that all minerals mentioned can be found. However, increasing amounts of typical Maas or Rhine minerals reflect increasing or decreasing reworking or supply of either Rhine or Maas sediments. The Lateglacial Maas and Niers terrace sediments differ from older sediments in a higher percentage of epidote, garnet and stable minerals, while the amount of staurolite and tourmaline is lower (Table 5.4). This can be used to distinguish Lateglacial from older deposits.

sediment	n	G	E	A	H	C	V	S	U	T/S	M	T	O
Niers terraces													
Holocene floodplain	3	30	15	11	9	1	7	12	2	5	3	5	33
Gennep terrace	2	13	21	22	10	1	7	7	1	5	5	8	69
Milsbeek terrace	1	34	12	19	11	0	8	5	1	5	3	3	30
Niers-Maas confluence area													
Holocene floodplain	4	16	18	9	6	1	3	24	2	6	4	12	61
Wanssum terrace	2	19	15	10	8	2	15	13	1	3	8	9	54
Broekhuizen terrace	2	14	16	13	12	0	5	21	2	5	6	8	64
Vierlingsbeek terrace	15	23	15	16	8	1	8	11	3	4	4	8	43
Rijkevoort terrace	9	21	16	25	9	0	7	5	2	5	4	6	33
aeolian deposits													
Holocene dune	1	34	12	16	8	2	10	3	2	3	7	7	48
Younger Dryas riverdune	2	25	13	18	10	2	11	7	3	3	5	5	46
coversand	1	38	19	8	0	1	1	19	3	4	4	6	25
Maas terraces													
Holocene floodplain	6	15	14	6	6	2	4	26	1	7	5	13	62
Wanssum terrace	3	19	17	6	3	1	3	23	1	8	7	13	45
Broekhuizen terrace	7	12	14	6	2	1	5	32	1	8	5	13	66
Vierlingsbeek terrace	12	15	14	11	5	1	3	24	3	7	6	12	47
Rijkevoort terrace	29	25	14	13	6	1	4	19	1	5	5	9	33
aeolian deposits													
Holocene dune	1	25	14	9	1	0	1	15	1	13	6	15	75
Younger Dryas riverdune	3	26	10	10	2	0	1	28	1	7	3	12	40
coversand	4	29	15	9	3	0	2	19	2	7	5	10	30
Diverse													
Overloon terrace	2	12	9	8	4	0	2	9	1	13	6	28	37
Pre Weichselian deposits	3	12	11	16	5	2	4	16	2	10	6	17	33
local creek deposits	4	32	16	15	6	1	5	8	1	6	5		

Table 5.4. Average heavy mineral content in % per terrace or aeolian deposit in the Maas and Niers valleys and confluence area. n=number of samples; G=garnet group; E=epidote group; A=alterite group; H=hornblende group; C=chloritoid group; V=volcanic group; S=stable group; U=unstable group; TS=topaz/staurolite group; M=metamorphic group; T=tourmaline; O=opaque minerals.

5.4.1 Comparison of Maas and Niers sediments

The Niers sediments are dominated by Rhine derived minerals as shown by higher amounts of alterite, hornblende and volcanic minerals, while the Maas sediments have higher amounts of chloritoid, staurolite, metamorphic minerals and especially tourmaline and stable minerals (typical for Maas). An overall higher amount of opaque minerals is also seen in the Maas sediments.

The mineral composition in the Niers-Maas confluence area shows a higher amount of hornblende, alterite and volcanic minerals and a lower amount of tourmaline and stable minerals when compared with the Maas sediments upstream. This reflects the downstream influence of Rhine derived minerals which were transported by the Niers.

The amount of tourmaline, stable- and opaque minerals increases in the younger Maas terraces, while the amount of garnet and alterite decreases. The overall increase in Maas derived minerals and decrease in Rhine minerals from the Pleniglacial Rijkevoort terrace to the Holocene floodplain reflects the diminishing amount of reworking of older mixed Rhine - Maas sediment during the Lateglacial and the enrichment by fresh supplied Maas sediments. The Wanssum terrace sediments do not fit this trend (see the garnet, stable mineral and opaque content in Table 5.4) because a deep incision during the Younger Dryas (Huisink 1997) provided older, more Rhine derived sediments. The minerals in the Niers-Maas confluence area show the same trend as in the upstream Maas terraces : garnet and alterite decrease and tourmaline, stable and opaque minerals increase with decreasing age. This suggests that also downstream of the Niers confluence diminished reworking resulted in an increase of Maas minerals. The Wanssum terrace sediments deviate again from this trend, which is seen for instance in the increase of garnet, volcanic and stable minerals.

5.4.2 Aeolian deposits

All aeolian samples show high amounts of garnet, which is in accordance with other studies (Crommelin 1964 and Vandenberghe and Krook 1981). The coversand samples found on top of the Rijkevoort terrace up- and downstream of the Niers-Maas confluence area show a similar heavy mineral content which suggests a similar source area for these aeolian sands. A gradual transition from fluvial via fluvio-aeolian into aeolian sands is observed in the upper part of the Rijkevoort terrace (Huisink 1997) and is reflected in the high amount of garnet in the upper fluvial sediment of the Pleniglacial terraces in both the Maas and Niers valleys.

The large parabolic dunes on the east bank of the Maas originated during the Younger Dryas (Bohncke et al. 1993). The source area for these dunes must have been the Younger Dryas floodplain as the mineral composition and grain size characteristics (Huisink 1997) of the Wanssum terrace sediments compare well with the riverdune sands. This confirms the reconstructed westerly winds during this period (Isarin et al. 1997). Samples from the river dunes were collected up- and downstream of the Maas-Niers confluence. The river dunes in the confluence area of the Niers and Maas show distinctly higher amounts of Rhine-derived minerals like alterite, hornblende and volcanic minerals, while the amount of stable minerals and tourmaline is much lower. This is in accordance with the different mineral composition of the Wanssum terrace sediments downstream of the Maas-Niers confluence. The Holocene aeolian dune sand was most probably reworked from the Younger Dryas river dunes and the mineral content is therefore similar to that of the river dunes.

5.4.3 Comparison with up- and downstream valley reaches

Comparison with other sediment petrographical studies appeared to be difficult as most heavy mineral studies in the Maas and Rhine valleys involved older fluvial sediments (Zonneveld 1956, 1974, De Jong 1956, Krook 1993). A petrographical study of Lateglacial and Holocene sediments was performed downstream in the "Land van Maas en Waal" by Hoek and Schorn (1990). During the Lateglacial, parts of this area were used simultaneously by both the Rhine and Maas, which resulted in mixed mineral assemblages. The Maas influence increased during the Younger Dryas and Holocene in the "Land van Maas en Waal" (Hoek and Schorn 1990). According to Huisink (1997) abandonment of the Niers valley by the Rhine took place probably in the Late-Bølling after the formation of the Gennep terrace. Since then the fresh supply of Rhine sediments stopped which might explain the increased Maas influence on the minerals downstream during the later part of the Lateglacial.

For a comparison with the upstream Maas valley petrographical reports of the former State Geological Survey (RGD) were used. The heavy mineral analysis of the RGD have been regrouped in this study to enable a comparison (Appendix A, Table 5.3). The average heavy mineral composition is shown in table 5.5. With the help of the coordinates and surface altitudes the RGD- samples could be attributed to the different Lateglacial terraces of Fig. 5.1. In general the two data sets (tables 4.5 and 5.5) correlate quite good, apart from some small differences related to defining epidote from alterite and grouping minerals to the hornblende or volcanic minerals groups.

Sediment	n	G	E	A	H	C	V	S	U	T/S	M	T	O
Niers-Maas confluence area													
Wanssum terrace	1	18	23	5	10	2	7	19	0	5	4	7	24
Vierlingsbeek terrace	14	17	16	21	7	0	13	6	0	6	4	9	33
Rijkevoort terrace	16	21	14	27	9	0	12	4	0	4	3	6	12
Maas terraces													
Holocene floodplain	5	17	7	10	10	0	3	29	0	9	4	11	
Wanssum terrace	14	12	11	12	6	1	3	12	0	13	9	22	
Broekhuizen terrace	10	16	9	8	4	3	1	21	0	12	7	20	
Vierlingsbeek terrace	21	12	17	13	8	0	1	19	0	10	7	13	
Rijkevoort terrace	40	19	13	18	6	0	9	8	0	8	5	13	
Overloon terrace	2	36	11	12	11	0	0	18	0	4	2	9	

Table 5.5. Average heavy mineral composition of samples analyzed by the former State Geological Survey in percentages, regrouped. Legend see Table 5.4.

In the upstream Lateglacial Maas terraces a trend is present in the heavy mineral composition of the sediments. The Maas derived minerals increase in younger terrace deposits as reflected by the increase of stable minerals and a decrease of epidote and alterite. This trend is similar to the trend observed downstream in my study area. The heavy mineral composition of the Wanssum terrace does not fit in this trend, related to the incision in the Younger Dryas and mixing with older Rhine-Maas sediments. The sediments downstream of the Niers Maas confluence show a higher amount of epidote, alterite and volcanic minerals and a lower amount of stable minerals, topaz/staurolite and tourmaline. This is again similar to my data set and it reflects the larger influence of Rhine-derived minerals in the confluence area of Niers and Maas.

The study of bulk and clay geochemistry of Maas sediments by Tebbens et al. (1998) shows increasing amounts of soil derived clay minerals during the Lateglacial and Holocene, with the exception of the Younger Dryas period. This deviation from a Lateglacial trend is comparable to that seen in the heavy mineral analysis and is, amongst other factors, also explained by increased reworking in the Younger Dryas.

5.5 CONCLUSIONS

The Lateglacial fluvial style changes and associated phases of incision and deposition are well reflected in the sediment petrography; especially in the heavy mineral composition. The gravel analysis proved to be less detailed and useful, but supported the results form the heavy mineral study.

The Niers terraces differ from the Maas terraces in the Niers-Maas confluence area in a higher greygreen sandstone and a slightly higher lydite content, and a lower milky quartz and total quartz content. The Niers served as a Rhine branch during the Pleniglacial and Bølling which resulted in more Rhine derived gravels in the Niers terrace sediments. The lower amount of quartz seems typical for Rhine sediments. The higher amounts of alterite, hornblende and volcanic minerals in the heavy mineral analysis of Niers sediments confirms the Rhine influence in the heavy minerals as well.

The Maas terraces upstream of the Niers confluence show a mixed mineral composition since both Rhine- and Maas derived sediments are found. Compared with the Niers terraces the Maas terrace sediments clearly show a lower amount of Rhine derived minerals and a larger influence of Maas derived minerals such as tourmaline and stable minerals. In the confluence area of the Niers and Maas a mixed heavy mineral assemblage was formed which is intermediate between the mineral composition of the sediments in the Maas and Niers valleys upstream of the confluence.

During the Lateglacial an increasing supply of Maas minerals is observed with decreasing age in the successive terraces, both up-and downstream of the Niers confluence. The increased Maas influence is shown by the increase of tourmaline, stable- and opaque minerals and a decrease in garnet and alterite. This trend is confirmed by the gravel analysis. This gradual change in sediment petrographical composition is explained by the diminished reworking of older, mixed Rhine-Maas sediment, related to the fluvial style changes during the Lateglacial. The change from a braiding river via a transitional stage to a meandering river resulted in a decreased reworking of sediment. The increasing influx of Maas minerals and gravels during the Lateglacial is not observed in the Younger Dryas sediments as a result of the deep incision during this period whereby older, Rhine-derived sediments were reworked. The fluvial style changes and associated phases of incision and deposition during the Lateglacial are thus reflected in the petrographical composition of the sediment.

It is demonstrated that the source area for the Younger Dryas river dunes on the east banks of the Maas was the Younger Dryas floodplain as the heavy mineral composition of the Wanssum terrace (Younger Dryas age) matches that of the Younger Dryas dunes, both up- and downstream of the Maas-Niers confluence.

6 Changing river styles in response to Weichselian climate changes in the eastern Netherlands

by M. Huisink. Submitted to Sedimentary Geology

6.1 ABSTRACT

In the Vecht valley, eastern Netherlands, distinct changes in fluvial style and erosional phases took place from the middle part (Pleniglacial) of the last glacial to the Holocene. The river changed from a low energetic laterally inactive river in the Weichselian Middle Pleniglacial into a braided river in the Late Pleniglacial. Furthermore a change from fluvial to aeolian dominated sedimentation occurred during the Late Pleniglacial. In the Lateglacial the river incised and became a low-sinuosity meandering river at first and later on a braided river in the Younger Dryas period. These fluvial changes are accompanied by erosional phases and both are most likely the result of climatically induced changes in the water - sediment discharge ratio. Vegetation appears to be an important factor as it determined bank and soil stability and therefore sediment supply but also evapotranspiration rates, which influenced discharges characteristics. A model explaining the fluvial response to climate change is presented and discussed by the example from the Vecht river.

6.2 INTRODUCTION

Geologists and geographers have become increasingly interested in the responses of fluvial processes to changed environmental conditions. Dutch fluvial studies in the eighties concentrated mainly on fluvial architecture and Late Pleistocene stratigraphy (Van Huissteden et al. 1986, Vandenberghe et al. 1984, 1987). Later on fluvial style changes were linked to changes in climate (Bohncke et al. 1993, Kasse 1995, Kasse et al. 1995a, Vandenberghe et al. 1994, Berendsen et al. 1995) or to tectonic movements (Van den Berg 1996). Vandenberghe (1993) proposed a non-linear model explaining periglacial fluvial style changes by changes in climate. To prove the general applicability, the model has to be tested for multiple climate changes and more regional settings. Validations took place so far in the Maas Valley for Lateglacial climate changes (Huisink 1997) and in Eastern Germany (Mol 1997). Further validation is needed however, since the fluvial responses to climate changes or local phenomena is still not completely understood. In the present morphological and sedimentological study the non-linear model will be tested and refined by concentrating on a small river valley in the Eastern Netherlands, the Overijsselsche Vecht (Fig. 6.1), and its behaviour to climate changes from the Middle to Late-Pleniglacial- (ca. 27 ka), Late Pleniglacial to Lateglacial (13 ka) and Lateglacial to Holocene (10 ka) transitions (Table 6.1). Weichselian climatic and environmental reconstructions have become increasingly detailed and well dated (Renssen 1997, Isarin 1997, Hoek 1997a,b, Huijzer and Vandenberghe 1998), which provides a good palaeoclimatic framework for this study.

The Vecht is a small, rainfed river with a catchment of 3.700 km^2 which originates in Germany and enters The Netherlands in the east (Fig. 6.1). The palaeo-Vecht valley was formed during the Saalian, when a continental ice sheet covered the northern part of the country and meltwater scoured a wide valley in front of the ice (Ter Wee, 1966). The Overijsselsche Vecht became confined to a narrow floodplain in the Lateglacial. Remnants of Lateglacial terraces can still be found (Fig. 6.1) although the terraces were eroded heavily in the Holocene by lateral migration of the highly sinuous river. The Pleniglacial evolution of the Vecht is probably comparable that of the Dinkel, which was studied in detail by Van der Hammen and Wijmstra (1971), Van Huissteden et al. (1986), Van Huissteden and Vandenberghe (1988), Van Huissteden (1990) and Ran (1990) (Table 6.1). The palaeo-Dinkel valley was filled with glacio-fluvial and lacustrine sediments in the Saalian, and with fluvial deposits and peat during the Eemian and early Weichselian. The early Pleniglacial is mainly characterised by erosion (Van Huissteden and Vandenberghe 1988), while aggradation by low energetic rivers occurred in the Middle-Pleniglacial. The Late Pleniglacial is dominated by higher energetic river sedimentation at first which changed during the later part of the Late Pleniglacial into aeolian sedimentation (Van Huissteden and Vandenberghe 1988). The sediments from this study are correlated with the stratigraphy of the Dinkel valley in Table 6.1.

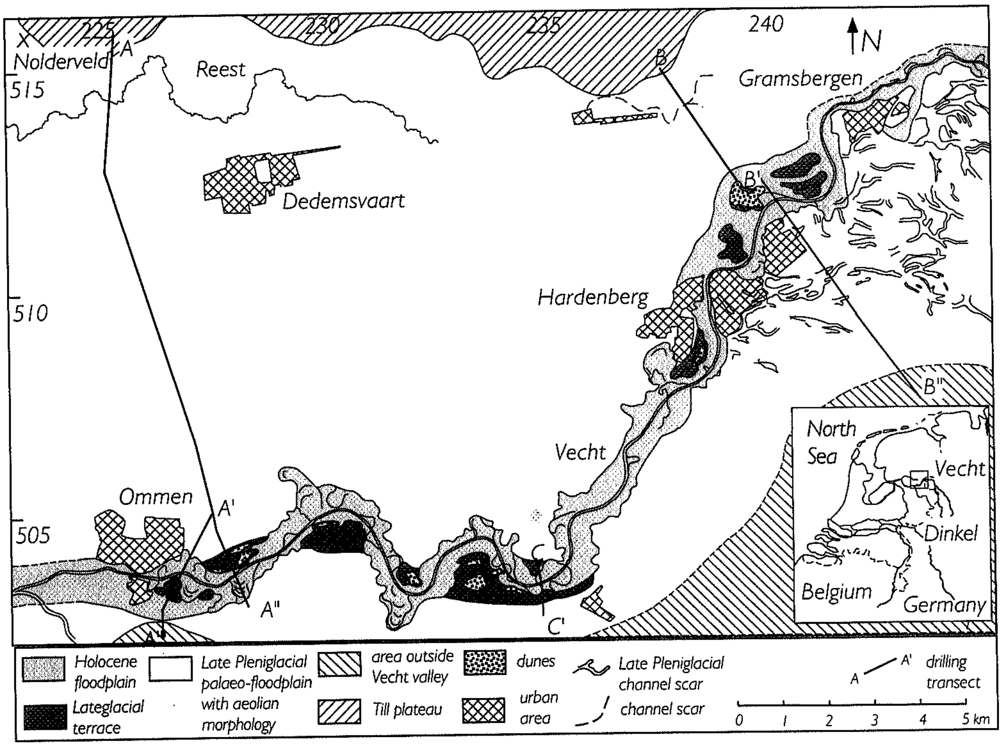


Figure 6.1 Location map of the Vecht valley with position of terraces.

Several methods were used to reconstruct the former river activity. Apart from up to 7 m deep hand drillings, arranged in transects across the valley (Fig. 6.1) two, up to 14 m deep mechanical drillings provided undisturbed sediment cores which are used to describe the sedimentary structures and to determine the depth of the Weichselian deposits in the Vecht valley. Ground penetrating radar has been used to track down channel fills and estimate channel dimensions. Exposure Nolderveld enabled a detailed sedimentary analysis of the top of the Late Pleniglacial deposits. Heavy mineral analysis were used to determine changes in source area. From the fraction 150-420 μm 100 non-opaque grains were counted and divided in the following groups: garnet, epidote, alterite, hornblende, volcanic minerals, stable minerals, staurolite, metamorphic minerals and tourmaline. Radiocarbon dates were used to date the units.

age ka	Chronostratigraphy			$\delta^{18}O$	lithostratigraphy by Van Huissteden and Vandenberghe 1988		lithostratigraphy by Van der Hammen 1971		facies and units by Huisink (this paper)
	Holocene				Singraven Formation		Singraven Formation		unit 6 fluvial facies a, b
10	Weichselian	Lateglacial	Younger Dryas	1	Twente Formation		Twenthe Formation	Wierden Member	unit 5 fluvial facies a, b, c, d aeolian facies e
11	Weichselian	Lateglacial	Allerød		Twente Formation		Twenthe Formation	Wierden Member	unit 5
12	Weichselian	Lateglacial	Older Dryas		Twente Formation		Twenthe Formation	Wierden Member	unit 5
	Weichselian	Lateglacial	Bølling		Twente Formation	Wierden Member	Twenthe Formation	Wierden Member	unit 5
13	Weichselian	Pleniglacial	Late	2	Twente Formation	Lutterzand Member	Twenthe Formation	Lutterzand Member	unit 4: aeolian facies f, g
	Weichselian	Pleniglacial	Late	2	Twente Formation	Beverborg Member	Twenthe Formation	Lutterzand Member	unit 4: aeolian facies e fluvial facies a, b
27	Weichselian	Pleniglacial	Middle	3	Twente Formation	Tilligte Member	Twenthe Formation	Mekkelhorst Member	unit 3 fluvial facies a, b, c
59	Weichselian	Pleniglacial	Early	4	Twente Formation	Dinkel Member	Twenthe Formation	Dinkel Member	
74	Weichselian	Earlyglacial		5a-d	Twente Formation	Liendert Member	Twenthe Formation	Liendert Member	
111	Eemian			5e	Asten Formation		Asten Formation		unit 2 fluvial facies c

Table 6.1. Weichselian chrono- and lithostratigraphy of the Vecht area.

6.3 RECONSTRUCTION OF SEDIMENTARY ENVIRONMENTS THROUGH TIME

The sedimentary facies (Fig. 6.2, 6.3) are described and interpreted according to Van Huissteden (1990), who based his classification on core-data and exposures. Most facies classifications are based on stratification for which exposures are required (Cant and Walker 1976, 1978, Cant 1978, Rust 1978b, Miall 1978, Williams and Rust 1969). Bridge (1985) argued that interpretation of palaeo channel patterns from ancient alluvium should require three-dimensional, large outcrops in which completely preserved sections of channels belts and overbank deposits are present. These requirements are not met in this area, but by combining sedimentological information from drillings and channel scar morphology, georadar images and shallow exposures the changing depositional environments have been reconstructed.

6.3.1 Sedimentation before the Middle Pleniglacial ($^{\delta}$O stage 3)

The oldest sediments of unit 1 (facies 1f and 1g, transect B-B'-B" and A-A'-A") are located outside the palaeo-Vecht valley (Fig. 6.1). Facies f consists of a mixture of clay, sand and gravel and was deposited by ice during the Saalian, forming the till plateau north of the Vecht valley. Facies g consists mostly of sands which were deformed by ice during the Saalian. Unit 2, consisting of sand and clay (transect A-A'-A" Fig. 6.2) is the oldest unit within the palaeo-Vecht valley. Pollen analysis from humic clay show a fresh-water vegetation of a warm-temperate climate (Bohncke pers. comm.), which most probably indicates Eemian deposition (oxygen isotope stage 5e).

sample	GrN	coordinates	depth	C14-age BP	material	interpretation + facies
A4-3	18180	225.800/502.600	3.75	35.700^{+3800}_{-2600}	peat	Middle-Pleniglacial, 3b
A37-3	18181	224.680/513.450	4.75	8.000 "60	peat	Holocene, 6a
A44-7	18182	227.750/503.300	6.55	21.000^{+1000}_{-900}	plant remains	mix Holocene and Middle Pleniglacial
B5-4	18183	238.100/511.225	4.00	31.500^{+3600}_{-2500}	humic clay	Middle Pleniglacial, 3b
YB4-7	18184	233.825/509.000	6.03	11.320 " 80	peat	Lateglacial, 5c
YB11-3	18185	234.300/502.925	2.70	31.370^{+700}_{-640}	peat	Middle Pleniglacial, 3c
ZB37-4	18186	238.100/514.725	5.05	31.400^{+1800}_{-1500}	peat	Middle-Pleniglacial, 3c
ZB44-6 >180 mu	18453	240.400/511.725	3.20	42.430^{+1060}_{-640}	peat	Middle Pleniglacial, 3c
ZB44-6 <180 mu	22124	240.400/511.725	3.20	40.820^{+2400}_{-1800}	peat	Middle Pleniglacial, 3a
	GrA					
A53-8	3255	227.350/504.300	4.75	11.070 " 90	plant remains	Lateglacial, 5b
A47-7	5324	227.580/503.750	4.00	5165 " 55	plant remains	Holocene, 6a

Table 6.2 C14 dates of bulk (GrN) and AMS (GrA)-samples from the Vecht valley.

6.3.2 Middle Pleniglacial fluvial sedimentation (stage 3, ±59-27 ka BP)

The Middle Pleniglacial sediments (unit 3) have a wide occurrence in the Vecht valley under 3 to 6 m of younger deposits (Fig. 6.2 and 6.3) and are characterised by an alternation of sands with peat or humic silts or clays that are frequently calcareous. Facies a (Fig. 6.2, 6.3) consists of medium to coarse (210-600 μm), moderately to well sorted sands, occasionally gravelly. A faint bedding or lamination may be visible, and finer-grained moderate to poorly sorted sand beds may be included.

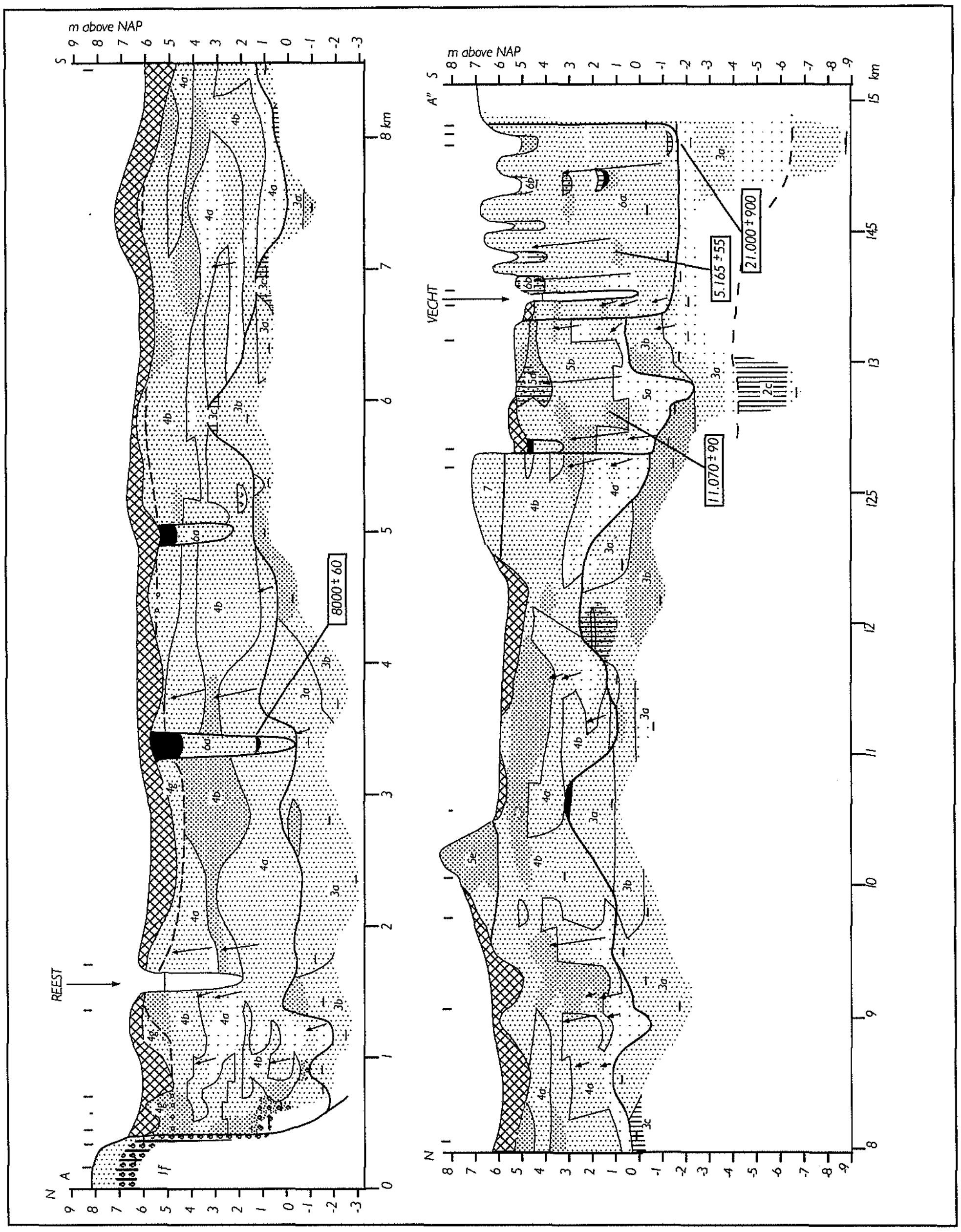


Figure 6.2. Cross section A-A'-A" of the Weichselian and Holocene valley fill. See for location and legend figures 6.1and 6.3.

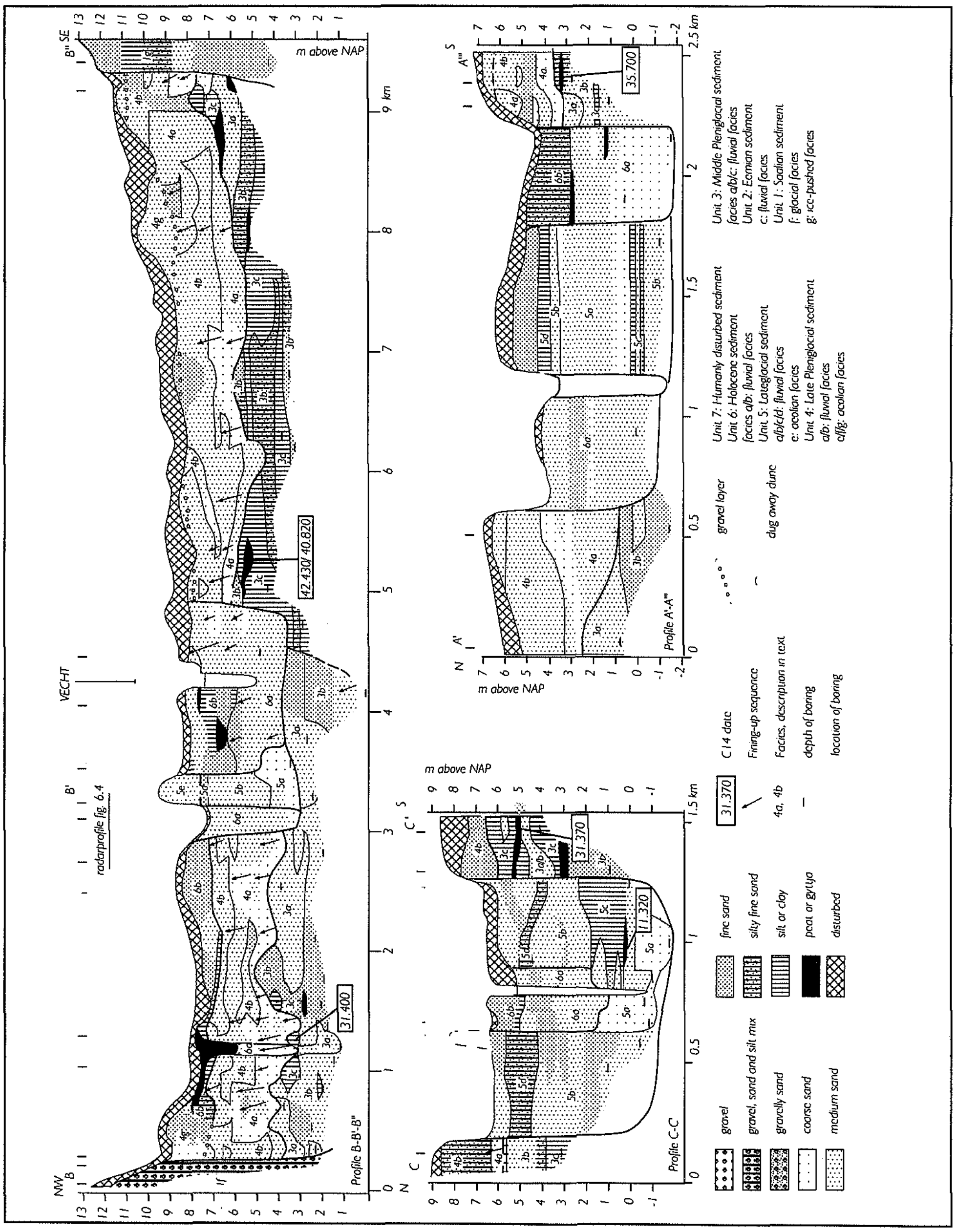


Figure 6.3. Cross sections B-B'-B''; C-C' and A'-A''' of the Weichselian and Holocene valley fill. For location see figure 6.1.

Fining-up sequences of up to 0.5 m occur sporadically. Facies b (Fig. 6.2, 6.3) consists of fine to medium (105-210 μm), mostly rather well sorted sand in which plant remains, clay pebbles and humic bands occur. Facies c consists of silt or clay alternating with organic layers. The organic layers consist of peat or very humic sand. Facies c is found mostly in the southern part of the area (profile B'-B" Fig. 6.3) where the humic clay, silt and peat layers are most abundant and up to 30 cm thick.

The deposits of facies a are interpreted as channel sands or crevasse-splay sands, while facies b probably represents overbank sands and facies c flood basin fines (after Van Huissteden 1990). The deposition of the Middle Pleniglacial sediments occurred in two phases: before and after circa 41 ka. In profile B-B'-B" (Fig. 6.3) erosion of the oldest deposits can be recognized in the northern part of the valley. Thick floodplain deposits of older than 41-42 ka (Table 6.2) dominate in the southern part, while the northern part of the valley is dominated by channel and overbank deposits of younger age (± 31 ka) but at a lower topographic level. The other transects show a dominance of channel and sandy overbank sediments which range in age from 31 to 36 ka (Table 6.2, Fig. 6.2,3). This suggests a dominance of flood basin fines before 41 ka and a change to channel and overbank dominated deposition after 41 ka, separated by a phase of incision between 41 and 36 ka. This incisional phase is also recognized in the Dinkel valley by Van Huissteden (1990) dated between 38 and 40 ka.

The preservation of fine sands, silts, clays and thin organic layers indicates that the depositional environment before circa 40 ka was low-energetic. The absence of fining-up sequences and the occurrence of homogeneous sand bodies points to an absence of point bar-like deposits and rather fast deposition of sand bodies such as crevasse-splays. The river pattern can be classified as laterally inactive (following Nanson and Knighton 1996). A further subdivision into single-channel or anabranching or stable-sinuous or straight is hard to make because of the lack of morphological data.

More energetic conditions occurred after circa 40 ka. Again the absence of fining-up sequences and the occurrence of homogeneous sand bodies without clear channel lags indicate sheet-like deposition. The flood basin fines (3c) are found at the same altitudes as the channel and overbank sediments (3a/b) and appear to interfinger laterally. It seems as if the anastomosing, low-energetic fluvial system changed into a higher energetic, episodically active river which involved sheet-flow like sand deposition and locally fine grained flood basin deposition, possibly comparable with the ephemeral anastomosing river system described by Mol (1997). This river type can be distinguished from a more energetic braided system by its well developed flood basin fines and the absence of rapid lateral and vertical lithological changes. The term ephemeral anastomosing river might however be confusing as the anastomosing is used mostly for fine-grained, low-energy rivers (Nanson and Knighton 1996). The river pattern is also laterally inactive, but higher energetic than the river before circa 40 ka. The term ephemeral anstomosing river might however beconfusing since anastomosing is used mostly for fine-grained, low-energy rivers (Nanson and Knighton 1996). This river can be typified as an intermediate energetic (intermediate between the former, low energetic river and a highly energetic braided river) laterally inactive river.

6.3.3 Late Pleniglacial fluvial sedimentation (stage 2, ± 27-13 ka BP)

The fluvial sediments (facies a and b) are often covered by aeolian sands (facies e,f,g Fig. 6.2, 6.3), which comprise together unit 4. During the Holocene, the valley became covered by peat, which was removed by man during the last centuries. This resulted, together with the presence of aeolian cover sands in a rather flat to gently undulating area. East of Hardenberg (Fig. 6.1) small anabranching channel remnants are found, orientated in a mostly east-west direction.

The major difference of unit 4 compared to the Middle Pleniglacial deposits is the absence of silty, clayey and organic sediments. The deposits are always non-calcareous and mostly moderately sorted. Facies a consists of medium to coarse (210 to 1400 μm), slightly gravelly sands. Occasionally, up to 4 cm gravels and clay pebbles are found. The vertical sequence is characterized by abrupt changes in grainsize and fining-up sequences of up to 1.5 m. In facies b, consisting of finer grained sand (75 - 210 μm, without gravels), fining-up sequences are usually absent, while lithological changes are more gradual.

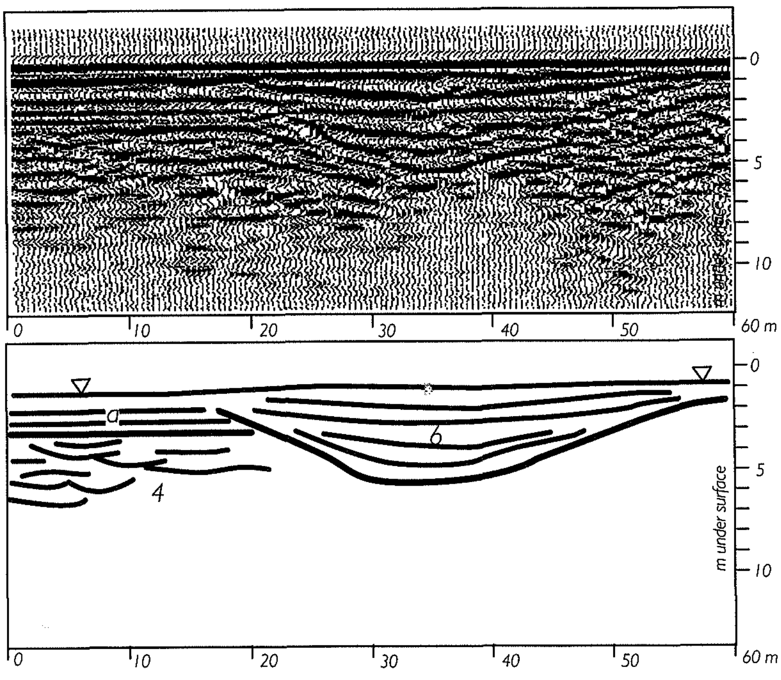


Figure 6.4 Georadar images showing the large dimension of a Holocene channel (facies 6) versus the smaller dimensions of Late Pleniglacial channels (facies 4) and the occurrence of cover sands (facies a).For location see figure 6.3.

The absence of silty, clayey or organic sediments, the occurrence of rapid vertical lithological changes, stacked short fining-up sequences and the pattern of multiple, anabranching channels indicate deposition in a highly energetic, braided river system. Within a braided river silty, fine-grained deposits (facies b) are found in abandoned channels or topographic lows while coarse-grained, gravelly sand or gravels (facies a) are found in channel - bar complexes (Williams and Rust 1969). Channel lags can be recognized in facies a, as thin gravel layers or very coarse gravelly sand layers. Facies a can be compared with lithofacies types St and Sp of Miall (1978) which represent dunes, bars and sand waves. Facies b correlates with lithofacies Sl and Fl (Miall 1978), which are crevasse-splays, antidunes, overbank deposits and waning flood deposits. Facies a represents the higher energetic sediments within the braided river floodplain and it is found predominantly at the base of unit 4 and in the northern part of the palaeo-floodplain (Fig. 6.2, 6. 3). Facies b, representing deposits from the less energetic parts of the floodplain, interfingers with facies a and it is more dominant in the upper part of unit 4 and in the central part of the palaeo-valley. The braided channel pattern is illustrated in georadar image Fig. 6.4 by a rather wild pattern of reflectors in which trough cross bedding of multiple small shallow channels is visible.

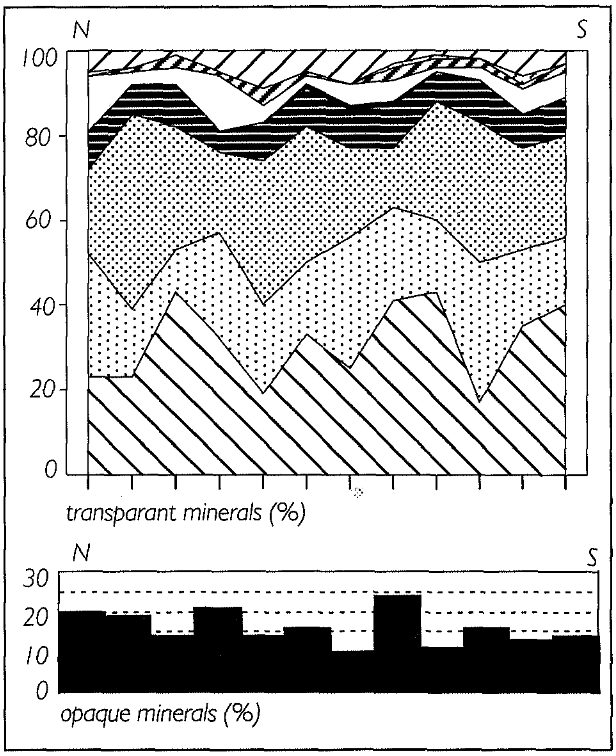


Figure 6.5 Heavy mineral composition of the Late Pleniglacial fluvial deposits in the Vecht valley; arranged on a north-south transect (samples from cross section A-A'-A" of figure 6.1), for legend see figure 6.6.

The braided river system occupied the whole valley at first, but concentrated into channel belts during the Late-Pleniglacial (Fig. 6.2, 6.3), which could indicate a change into less energetic conditions. The position of the Late-Pleniglacial channel belts possibly determined the location of Lateglacial and Holocene channels as these are found at similar locations in the valley. Samples taken from the Late-Pleniglacial sediments on a north-south transect (transect A-A') to study petrographical changes within the Late-Pleniglacial palaeo valley, showed no significant variation in heavy mineral composition (Fig. 6.5). This indicates that the sediment was distributed evenly throughout the palaeo-valley which supports the origin of unit 4. Furthermore, the Late Pleniglacial deposits show a higher amount of garnet compared with the other Weichselian and Holocene deposits (Fig. 6.6) which might be related to an admixture of aeolian material in the fluvial sands since aeolian cover sands have a high garnet content (chapter 5). The Late Pleniglacial fluvial sediments are dated stratigraphically as they are situated on top of Middle Pleniglacial- and under Lateglacial deposits (Fig. 6.2, 6.3).

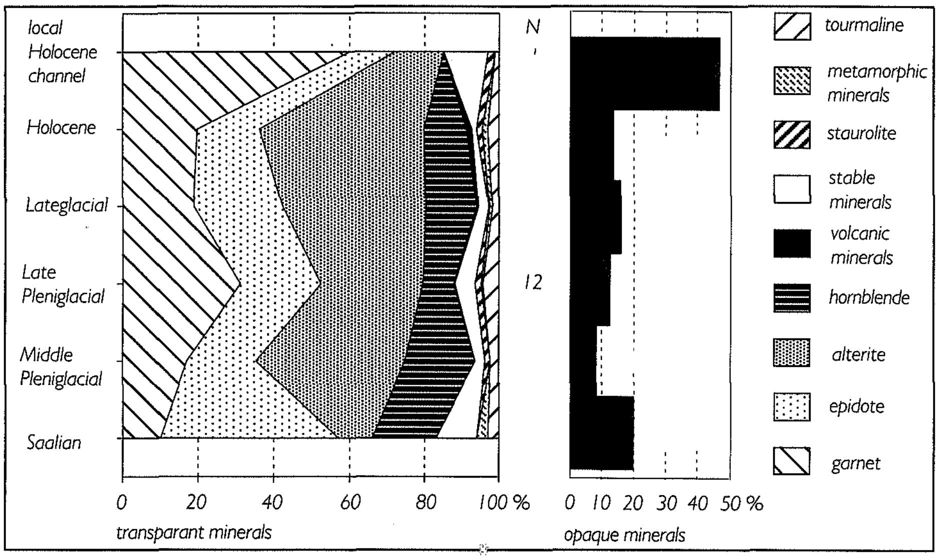


Figure 6.6 Average heavy mineral composition of Weichselian and Holocene fluvial sediments in the Vecht valley. N is number of samples.

6.3.4 Change from fluvial to aeolian sedimentation during the Late-Pleniglacial (stage 2, ± 18-13 ka BP)

In the most north-western part of the study area near Nolderveld (Fig. 6.1, Dutch coordinates 222.575 - 516.975), an exposure in the upper two metres of the valley sediments shows fluvial sediments at the base that change towards aeolian sands at the top (Fig. 6.7). The lowermost facies 4b, consists of moderately sorted and rounded sand and silt with occasionally coarser grained layers or lenses (bottom left of lacquer peel Fig. 6.7). Sometimes small-scale trough cross bedding and short fining-up sequences are visible, indicating fluvial activity. Most of the sediment however, is heavily cryoturbated (Fig. 6.7, 8). Within the fine sand vertical platy structures can be distinguished which are formed by thermal contraction in permafrost (Mol et al. 1993).

Facies 4e is found on facies b and consists of an alternation of fine sand and loamy fine sand with crinkly lamination and adhesion ripples in the top of the facies (Fig. 6.7, 6.8). This indicates deposition by wind on an alternating moist and dry surface (Schwan 1986, 1988, Kasse 1997). The grain size of facies 4e shows a distribution with a minor second peak in the silt fraction (Fig. 6.9, wet aeolian), which is typical for this facies (Schwan 1986, Vandenberghe and Van Huissteden 1989). The occurrence of multiple deflation layers, present as gravel strings in Fig. 6.7, is also typical for aeolian activity. Towards the base of the facies the distinction between the silty and sandy layers becomes less distinct and coarser grained lenses may be present, possibly related to shallow water flow. The top sediments of facies 4e are interpreted as "wet-aeolian" sands, while the sediments at the base are "fluvio-aeolian" sands that indicate the transition from fluvial to aeolian dominated deposition. The boundary between facies 4b and 4e is gradational which points to the gradual changes in sedimentary environment during the later part of the Late Pleniglacial.

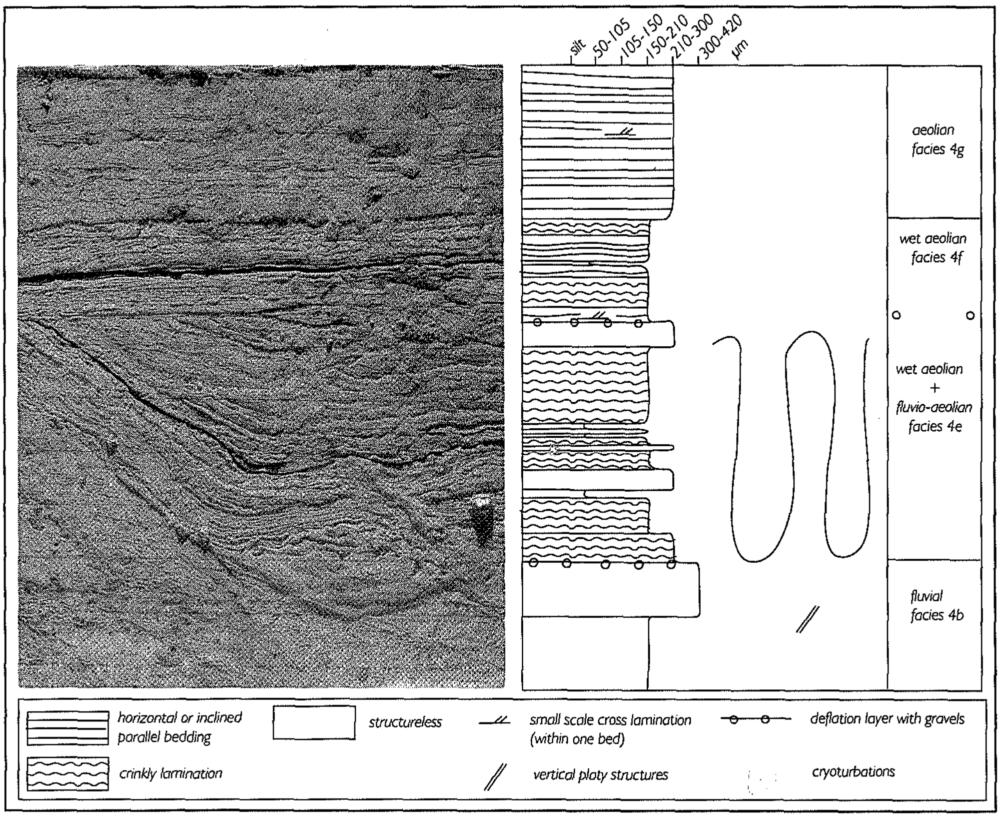


Figure 6.7 Laquer peel of the Nolderveld exposure showing the sedimentary structures of the transition from fluvial sediment at the base towards "dry"aeolian at the top. For location see figure 6.1.

Figure 6.8 Large-scale involutions in fluvio-aeolian sands (facies 4b) at Nolderveld, for location see figure 6.1.

Facies f is located above the Beuningen gravel bed and is not cryoturbated. This facies is very similar to the top of facies 4e (Fig. 6.7). Apart from crinkly lamination some parallel bedding is also visible (Fig. 6.7), which indicates that the surface became drier. This facies is also interpreted as "wet aeolian". Facies 4g consists of fine, well sorted sand without silty or coarse-grained laminae and it is generally finer grained than the underlying two facies (Fig. 6.9 aeolian). Horizontal or inclined parallel bedding dominates (Fig. 6.7) whic is shown nicely by horizontally parallel georadar reflections (Fig. 6.4). This sediment was deposited on a dry surface with low relief possibly by strong winds (Schwan 1987). Facies 4g correlates with the extensive sand sheet deposits that accumulated during the Late Pleniglacial and early Lateglacial in North-West and Central Europe (Kasse 1997).

Several deflation layers occur in facies e (Fig. 6.7, 6.8) which comprise the Beuningen Gravel Complex. The upper deflation layer is the Beuningen Gravel Bed which is used as a stratigraphic marker in The Netherlands as it usually marks the boundary between the fluvio-aeolian sands of the Beverborg Member and the aeolian sands of the Lutterzand Member. The Beuningen Gravel Bed at Nolderveld generally marks the transition between dry and wet aeolian deposits but it is also found under facies f (Fig. 6.7), which indicates that deposition of wet-aeolian sands occurred also after the formation of the Beuningen Gravel Bed. The Beuningen Gravel Bed is clearly erosive and truncates the underlying deposits which were subject to large-scale cryoturbation (Fig. 6.7, 6.8). Multiple phases of cryoturbation are recognized by involutions, ranging from 20 cm to 1.20 m in depth. The large (1.2 m) amplitude of the involutions (Fig. 6.7, 6.8) indicates progressive degradation of permafrost (Vandenberghe 1988) and a mean annual air temperature of ≤ -4° C (Vandenberghe and Pissart 1993).

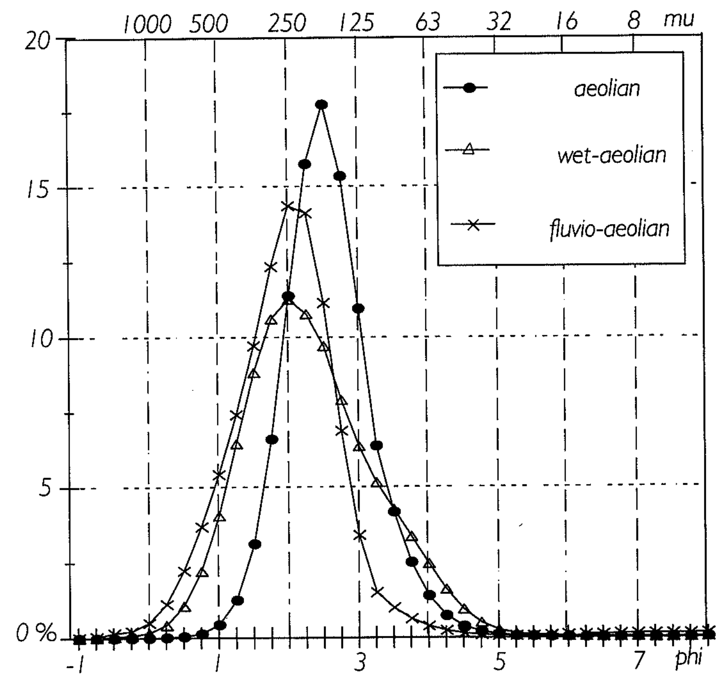


Figure 6.9 Grain size frequency curves of three samples from aeolian, wet-aeolian and fluvio-aeolian sands from Nolderveld. For location, see figure 6.1.

6.3.5 Lateglacial fluvial sedimentation (13-10 ka BP)

The Lateglacial palaeo-floodplain is located in the southern part of the Vecht valley and is much narrower than the previous, Late Pleniglacial floodplain (Fig. 6.1). Lateglacial terrace remnants are found along the present river at 1 to 2 m above the present floodplain. An estimated sinuosity of the Lateglacial palaeo-floodplain is 1 to 1.1, which is very low sinuous or straight. Channel scars on top of the terraces also show a straight morphology (Fig. 6.1).

The Lateglacial sediments (unit 5) are divided into four different facies. Facies a consists of coarse (300 - 2000 μm) grey sand which is sometimes slightly gravelly, moderately sorted and sometimes calcareous. A mechanical drilling on the Lateglacial terrace near Ommen provided undisturbed sediments in which large-scale cross-bedding is present, possibly related to large-scale trough cross-bedding. Facies b is fine to medium grained (up to 300 μm), rather well-sorted sand that may be humic or silty and is non- or slightly calcareous with many plant remains. Facies c is finer grained as it consists of clayey, sometimes humic silt, clay or peat. It is calcareous and contains also a lot of plant remains. Facies d consists of moderately sorted, very fine, silty sand or sandy silt, is sometimes humic and non-calcareous with locally loamy or humic pebbles.

Coarse channel sediments (facies a) are found at the base of unit 5, interfingering with finer grained channel fills, bars or levees (facies b). Facies c and d are floodplain sediments whereby facies c is deposited in standing water and facies d by flowing water. Two phases of deposition can be recognized. First, deposition by a generally low energetic, probably meandering river as well- developed flood basin fines (facies c, profile C-C') are found. Peat that formed part of the floodplain sediment was dated at 11.320 BP (profile C-C', Fig. 6.3, Table 6.2). Secondly, erosion of facies c took place (Fig. 6.2, profile A'-A'') by a higher energetic, probably braided river. The sediment characteristics change more rapidly in vertical and horizontal direction, channel sediments (facies a) occur more frequently and channel scars at the surface show a straight morphology. Plant remains dated at 11.070 BP were found in sediments from the braided river (profile A-A'). The plant remains consisted of a cold flora with species as Batrachium sp., Hippuris, Sparganium, Empetrum and Potentilla erecta which are known to occur frequently in the Younger Dryas period. This, together with the radiocarbon date suggests sedimentation of facies c during the early part of the Younger Dryas. Facies d is found in all profiles, quite close to the surface, topping facies a and b and locally interfingering with them (profile A'-A", Fig. 6.2).

Facies e of unit 5 comprises Lateglacial dunes, which are less well sorted and coarser grained than facies g of unit 4.The dunes are found on top of Lateglacial terraces. The absence of these dunes in the Holocene floodplain and the occurrence of a well-developed soil in them indicates a Lateglacial age. Since the dunes are found on top the Lateglacial terraces near Ommen in which the sediment was dated at 11.070 BP, the dunes are of Younger Dryas age.

6.3.6 Holocene fluvial sedimentation (10- 0 ka BP)

The Holocene fluvial sediments (unit 6) are found in the present floodplain (Fig. 6.1) and in solitary channel fills in the palaeo-valley (Fig. 6.2, 3). The Holocene Vecht was a rather narrow, single channel, meandering river with a sinuosity between 1.8 and 2.4. The River Vecht was canalized in the last two centuries, but before the canalisation it was underfit meandering within larger meander loops, with a sinuosity of 1.1 to 1.3. These large meander loops are possibly the result of the self-organising processes that occur within a meandering river and are more or less randomly formed (Stølum 1996). These self organising processes comprise cut-offs and lateral migrations which occur in ordered and chaotic states of the river and result in high-sinuosity meander loops that form a larger scale meandering planform (Fig. 6.1a in Stølum 1996).

The Holocene sediments differ from the Lateglacial sediments in grainsize (finer than 600 μm), but other characteristics are similar like colour, content of plant remains and occurrence of fining-up sequences. Facies 6a consists of up to 80 cm long, stacked fining-up sequences. Typical is the occurrence of reworked wood fragments. In a mechanical drilling large-scale cross bedding was observed that changed upwards into a repetition of sand with mud drapes (at least 21 cycles), representing falling water deposits after floods. The sediments are found in well preserved pointbars (Fig. 6.1) and clearly resemble pointbar sediments. Channel fills are also included in this facies. Facies b is finer grained, clayey or silty fine sand or sandy clay and silt. This facies is found in the swales between the pointbars (Fig. 6.2) and on top of facies a as floodplain sediment (Fig. 6.3). Facies c may be equivalent to facies d from unit 5 as both are floodplain sediments found at the same topographic level and with similar characteristics.

In georadar images like in Fig. 6.4 the solitary Holocene channels are recognized by a clear reflection of the channels base, possibly a gravel lag and the layered infilling of the channel. The location and depth of the channel in Fig. 6.4 was confirmed by drillings along transect B-B' (Fig. 6.3). The dimensions of these channels are much larger (5 m deep and 20 m in diameter) when compared with the braided channels of the Late Pleniglacial.

6.4 FLUVIAL RESPONSE TO CLIMATE CHANGE

Fluvial style changes can be explained as a reaction of a river to base level, tectonic or climatic changes. Sea level changes were unimportant as the sea was some 100 m lower than today (Jelgersma 1966). Evidence for tectonic activity during the Weichselian and Holocene was not found in the study area. The influence on river channel pattern of tectonic movements compared to climate changes is thought to be of minor importance as has been established in the Lateglacial Maas Valley (Huisink 1997). The catchment of the Vecht remained the same from the Middle Pleniglacial until the Holocene since the fluvial sediments show a comparable heavy mineral composition (Fig. 6.6). By eliminating sea level changes, tectonic activity or changes in source area, the changes in fluvial styles can, almost certainly, be linked to climatically induced changes in water and sediment supply ratios. Distinct changes in temperature occurred in the Weichselian as shown by the oxygen isotope curves of Greenland and Antarctica ice cores (e.g. Johnsen et al. 1995) and other proxies like periglacial features, vegetation and fauna (Table 6.1).

6.4.1 Fluvial response to cooling

The Late Pleniglacial (± 27-13 ka) marks the maximum cold phase of the Weichselian when ice sheets reached their maximum extent and continuous permafrost established in The Netherlands (Huijzer and Vandenberghe 1998). Shorter cooling phases took place from 41-38 ka (Hasselo Stadial after Ran and Van Huissteden 1990) and during the Younger Dryas (11-10 ka) when a transition from seasonally frozen ground to discontinuous permafrost occurred (Huijzer and Vandenberghe 1998).

The fluvial style changes from a low energetic river before circa 41 ka into a higher energetic, laterally inactive river after 36 ka and the associated incisional phase can most likely be explained by the climate cooling in the Hasselo Stadial. The establishment of discontinuous permafrost resulted in decreased evapotranspiration rates as more latent heat is needed to melt ice and less radiation is consequently used for evapotranspiration (see Woo and Winter 1993). The diminished evapotranspiration could have resulted in increased discharge versus sediment supply which led to erosion. The increasing cold could also have led to increasing river ice thickness which resulted in a higher dynamic river during the ice break up in spring. The resulting floods favoured deposition of sand in the floodplain, which together with a water level rise terminated the Middle Pleniglacial marshy vegetation growth in the floodplain (Ran 1990). As soon as the arctic to shrub tundra vegetation that occupied the river plains before (Ran and Van Huissteden 1990) became sufficiently destroyed bank and soil stability decreased, sediment supply increased and reworking and aggradation by a higher energetic river took place.

The transition from the Middle to Late Pleniglacial involved a major cooling event. The establishment of continuous permafrost and scarceness of vegetation resulted in increased peak discharges in a barren landscape where protection against erosion was minimal. A highly energetic braided river became active some time after 31 ka (Fig. 6.10). The base of the Late Pleniglacial braided river deposits is rather irregular as channels scoured locally at least 5 metres into the underlying deposits (Fig. 6.2, 3). Whether the change in fluvial style was accompanied by an incision phase is hard to establish since the Late Pleniglacial braided river used the entire valley. The change towards braided river conditions occurred in many European river valleys (Mol 1997) which suggests that the climate change was severe enough to cross internal thresholds in many rivers.

After 11.3 ka erosion and a change from a meandering to a braided river occurred. This is related to the cooling at the onset of the Younger Dryas which was distinct but short. In the previous Allerød period a rather closed, woody vegetation with birch and pine was present (Hoek 1997a). Evapotranspiration diminished as a result of the Younger Dryas cooling which led to increased peak discharges. Sediment supply was restricted, however, by the dense vegetation cover which resulted in erosion. As soon as the vegetation cover became more open and sediment supply increased a braided river was actively aggrading. The response of European rivers to the Younger Dryas cooling was diverse. Some rivers showed similar changes as the Vecht like in Britain (Rose 1995, Collins et al.1996); and the Ems and Niederrhein valleys in Germany (Klosterman 1995) and locally in Poland (Kalicki and Zernickaya 1995). Other rivers showed no erosion or change in morphology like the Warta in Poland (Vandenberghe et al. 1994) or the Somme in France (Levefre et al. 1995).

6.4.2 Fluvial response to increasing aridity

In exposure Nolderveld a gradual transition from fluvial to aeolian deposition took place in the Late Pleniglacial (Fig. 6.8). This has also been observed in the Maas valley (Mol et al. 1993, Huisink 1997). The increase of windblown sands in the floodplain is probably related to a severe aridity that followed the maximum ice-sheet advance after 20 ka (Schwan 1987). As the climate became warmer and the ice-sheet decayed the continuous permafrost melted which resulted in the large- scale involutions visible in Nolderveld. Kasse (1997) argued that the disappearance of permafrost resulted in an increased infiltration and enhanced soil water storage capacity which favoured deposition of sand sheet deposits, that are found throughout Northwestern Europe at that time.

6.4.3 Fluvial response to warming

The Lateglacial was warmer than the previous Pleniglacial period and probably also wetter as indicated by high lake levels (Bohncke and Wijmstra 1988). The Vecht incised and changed from a braided river into a low-energetic, low-sinuous meandering river, confined in a narrow floodplain. This can be explained by the better protection of the soil against erosion by the Lateglacial vegetation development, the more regular run-off pattern, the increase of evapotranspiration and larger soil water storage capacity. In contrast to climate cooling the maximum incision was not shortly after the climate change, but prograded because the vegetation development gradually diminished the sediment supply. A meandering river with a well-developed floodplain was active at and before 11.320 BP in the Vecht valley. Sediments from that period are found near the base of the Lateglacial sequence and it is therefore likely that the maximum incision in the Vecht valley occurred shortly before 11.3 ka. This gradual incision related to vegetation development was also found in the Maas valley (Huisink 1997). There the transition from the Late Pleniglacial braided river towards a meandering one was characterized by a transitional phase which showed characteristics of both braided and meandering rivers (Huisink 1997, Vandenberghe 1995).

The warming of the Holocene resulted in similar processes as compared to the warming of the Lateglacial. However, the re-establishment of a dense vegetation cover was much faster and the discharge regime was much more regular. The difference in hydrological characteristics probably explains the differences in magnitude and sinuosity between the meandering phases of the Lateglacial and the Holocene.

Incision and subsequent change in fluvial style towards higher energetic conditions took place after circa 41 ka and before circa 36 ka which is most probably related to the Hasselo Stadial (40 -39 ka) where a drop in the annual temperature of circa 5 degrees took place (Ran 1990) took place. Unfortunately a more exact dating of the incision was not possible, so debate is possible on the fluvial response to the Hasselo Stadial. One could argue that the incision took place as a response to the warming at the end of the Hasselo Stadial when temperatures rose to previous (warmer) values. This would not explain however, the higher energetic river in which more peaked discharges and sediment fluxes occurred. This is only expected when a more open landscape was present at that time, which was explained by the destruction of vegetation by increased floods, which in turn were related to diminished evapotranspiration or increased river ice thickness related to cooling.

6.5 MODEL FOR THE RESPONSE OF RIVERS TO CLIMATE CHANGES

The link between climate change and fluvial response is clear as distinct changes in climate were reflected simultaneously in the Vecht (Fig. 6.10) and other north-west European river valleys during the Weichselian. The interaction between discharge and sediment supply was the major control on fluvial style changes and erosional phases (conform Vandenberghe 1993). Discharge characteristics changed quickly when temperature, precipitation or permafrost conditions changed, whereas sediment supply lagged behind as this was closely related to vegetation which needed time to develop or disappear. At the beginning of a cold period evapotranspiration decreased, and soil water storage capcity decreased which resulted in higher peak discharges. Vegetation remained intact at first so bank and soil stability were maintained and the increased ratio of discharge versus sediment caused erosion. At the onset of a warm and wet period a pioneer vegetation stabilized the soil thereby decreasing the sediment load which resulted in incision. These processes occurred if local thresholds were exceeded and gradual transitions in fluvial style were explained as intrinsic evolutions (Vandenberghe 1995).

The Vecht responded to climate cooling according to the model of Vandenberghe (1993), but the responses of the Vecht to warmer and wetter conditions were more gradual than the response to cold conditions. The maximum depth of incision following the Lateglacial warming was reached shortly before 11.3 ka and not at the onset of the Lateglacial (around 13 ka). The fluvial response to the Younger Dryas cooling on the other hand was fast, as incision and a change towards braided conditions occurred between 11.3 and 11 ka. This gradual response to warming and fast response to cooling is explained by the direct response of the river to changed sediment to water discharge ratios. Besides the influence of the vegetation on bank and soil stability and thus sediment supply, it also influenced evapotranspiration rates and soil water storage

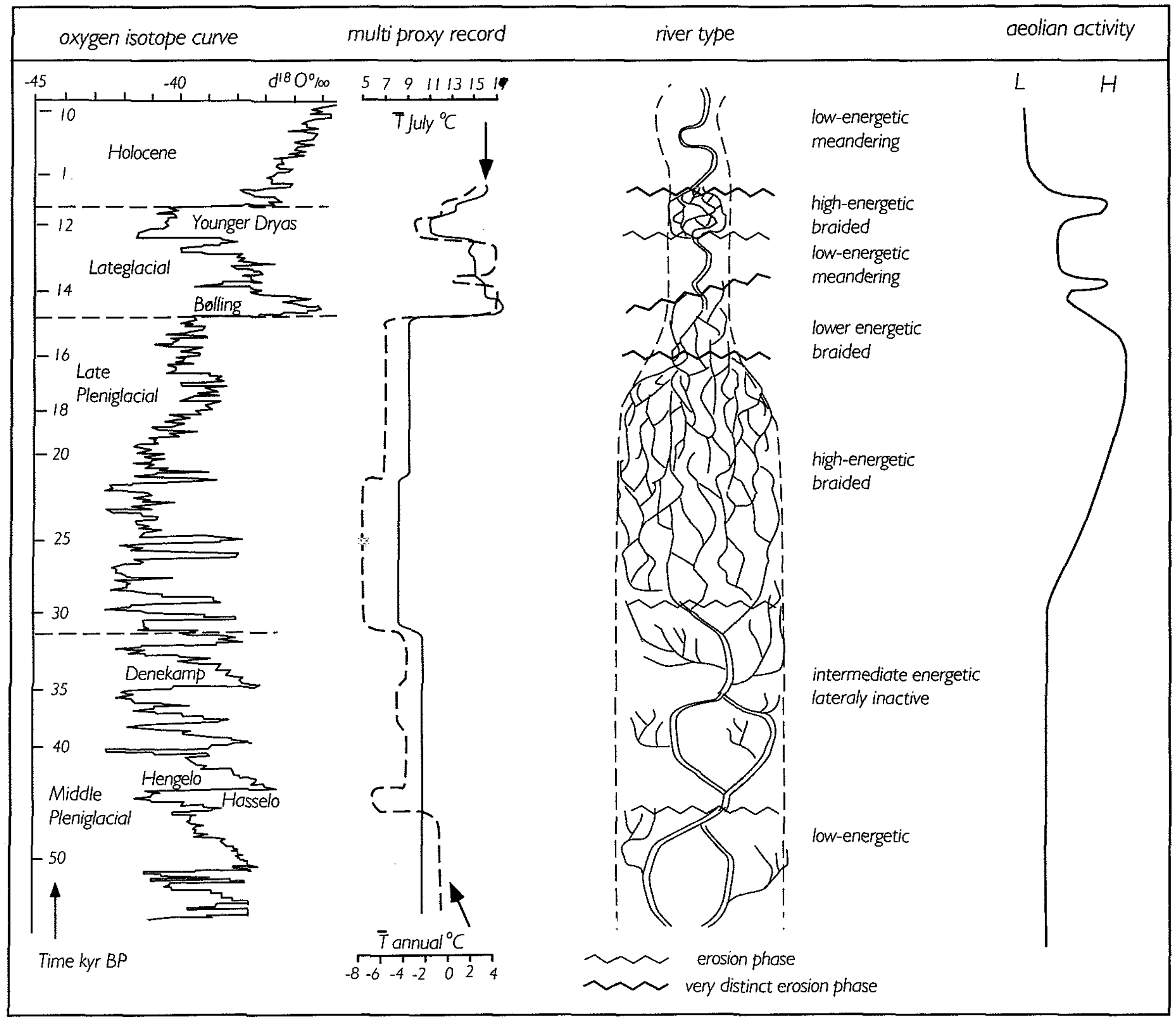


Figure 6.10 Compilation of fluvial changes in the Vecht valley from the Middle Pleniglacial to the Holocene with phases of erosion, aeolian activity, mean July and annual temperatures and oxygen isotope curve. Oxygen isotope curve after Dansgaard et al. (1993); Palaeo temperatures after Hoek (1997) and Huizer and Vandenberghe (1998).

capacity and therefore the discharge characteristics. The Younger Dryas cooling resulted in increased and more peaked discharges, while vegetation remained intact at first, so sediment supply remained limited. This caused a fast incision, within some 300 years in the Vecht valley. After a while the vegetation became destroyed and sediment supply increased and aggradation by a braided river occurred. The role of the vegetation cover was also very important at the onset of the Lateglacial, as the gradual development of the vegetation cover led to decreased sediment supply and gradual incision. The vegetation cover thus played an important role in the timing of erosion and aggradation.

Fluvial style changes and periods of incision occurred only when regional or local thresholds were crossed. The nature and magnitude of these thresholds remain obscure however, but variations in discharge characteristics and sediment load, determined largely by changes in vegetation cover, are important factors. All observed fluvial responses of the Vecht can be explained by changes in vegetation, which in turn were related to climate changes. The short warming and cooling events in the Middle Pleniglacial that were recorded in oxygen isotope curves did not result in significant vegetation changes (Ran 1990) and subsequently they did not lead to fluvial responses. On the other hand, all periods with major changes in vegetation (transition from Middle to Late Pleniglacial, onset Lateglacial, Younger Dryas period and onset of Holocene) coincided with distinct fluvial responses.

The cooling around 40 ka resulted in an erosional phase in both the Dinkel and Vecht valleys but the Dinkel did not change its fluvial style according to Van Huissteden (1990) as opposed to the Vecht. This suggests that the crossing of internal thresholds might also be related to the position in the river catchment. The Dinkel valley, as an upstream subcatchment, was perhaps too small to be seriously affected by the cooling event. The Vecht valley had a larger catchment and thus the impact of changes in water and sediment fluxes was larger.

6.6 CONCLUSIONS

The Overijsselsche Vecht reacted on Weichselian climate changes by changing its fluvial style repeatedly (see Fig. 6.10). The impact of changes in vegetation cover is large as this determined bank and soil stability and thus sediment supply into the river system. The vegetation cover also influenced evapotranspiration rates and soil water storage capacity and thus discharge characteristics. Incision took place rapidly at the onset of a cooler period when the discharge increased before the vegetation cover was destroyed. Incision also occurred as a response to warmer climate conditions but with a time-lag, depending on the time needed to establish a vegetation cover dense enough to decrease sediment supply and increase evapotranspiration rates. Climate changes that had an impact on vegetation resulted in a fluvial response in the Vecht valley. The transition from Middle to Late Pleniglacial, the onset of the Lateglacial, the Younger Dryas period and the onset of the Holocene were recognized in multiple river valleys by similar fluvial responses which stresses the importance of climate and vegetation in the crossing of thresholds by rivers. A comparison of the Vecht with the Dinkel showed a different response by both rivers to the cooling in the Hasselo Stadial. This indicates that catchment size and position in the catchment might also be factors that determined whether a change in fluvial style occurred or not. The short Middle Pleniglacial climate changes that did not affect vegetation were not followed by a fluvial response.

7 Synthesis

7.1 OBJECTIVES

Similarities in fluvial development in Northwestern and Central European lowland valleys suggest a strong forcing of climate change on rivers. The fluvial response to climate change is complex however and both local factors like geologic and tectonic setting and regional factors like climate change play a role. This study aimed to reconstruct and explain fluvial morphological and sedimentological changes from the mid-Weichselian until the Holocene. In this period four distinct climate changes took place. The complex and non-linear fluvial responses to these climate changes have been studied in the Maas and Overijsselsche Vecht valleys in The Netherlands. The valleys differ in size and geological setting in order to distinguish local from regional factors that triggered changes in geomorphological and sedimentological processes. To make a comparison between both areas the first question that needs to be answered is:

1. What are the changes in geomorphological and sedimentological processes from the Middle Pleniglacial to the Holocene in both study areas ?

In chapter 2 and 6 the fluvial developments of the Maas and Vecht have been described. In this chapter both rivers will be compared and several striking similarities in fluvial development between the two rivers have been observed but also some differences. The next question therefore is:

2. How can these fluvial changes in morphology and sedimentology be explained. Are they related to climate change or to other factors such as tectonic activity ?

In chapter 3 the tectonic and climatic influence on the fluvial development of the River Maas was determined by constructing detailed longitudinal terrace slopes. This provided information on tectonic movements in the Maas valley and the fluvial responses to those movements. Increasingly detailed Weichselian climatic and vegetational reconstructions have been published recently that enabled a correlation between fluvial changes and changes in temperature or aridity and vegetation development (chapters 2 and 6). Vandenberghe (1993) proposed a model to explain phases of erosion and deposition by climate and vegetation changes. In this synthesis this model will be evaluated and an answer to the following question will be formulated:

3. Do rivers react to climate changes according to the non-linear model proposed by Vandenberghe (1993) or are modifications or refinements to this model needed ?

Apart from these three main questions attention was paid to the following issues:

4. Is it possible to quantify the amounts of eroded, deposited or reworked sediment during a fixed period in order to indicate changes in the dynamics of the river in time ?
5. Are changes in amounts of erosion, deposition or reworking of sediment reflected in the sediment petrography ?

These questions are answered in chapters 4 and 5 and will also be summarized here.

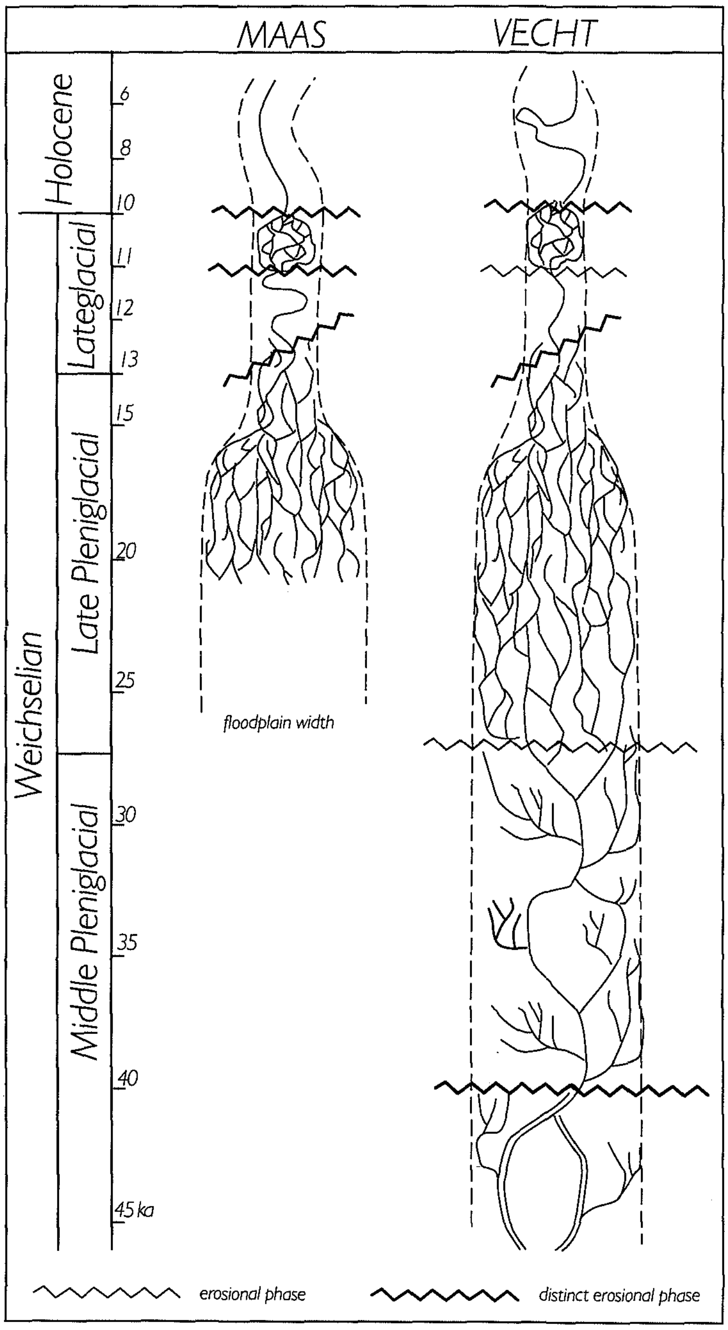


Figure 7.1 Reconstructed fluvial styles of the Maas and Vecht for the Pleni- and Lateglacial and Holocene.

7.2 CHANGES IN GEOMORPHOLOGICAL AND SEDIMENTOLOGICAL PROCESSES IN THE MAAS AND OVERIJSSELSCHE VECHT VALLEYS FROM THE MIDDLE PLENIGLACIAL UNTIL THE HOLOCENE

During the first part of the Middle Pleniglacial a low-energetic river was active in the Vecht valley which deposited thick floodplain sediments. The absence of point-bar deposits, the occurrence of crevasse-splay sands, and a low-energetic environment could possibly point to an anastomosed river system (Fig. 7.1). Between 41 and 36 ka a significant erosion took place and a change in fluvial style towards higher energetic conditions occurred (Fig. 7.1). This river was not as energetic as a braided river as floodplain fines are still present. Homogeneous sheet deposits dominate instead of heterogeneous channel and bar sediments which suggests that the river was laterally inactive. The fluvial system is possibly comparable with the sandy, ephemeral anastomosing system as described by Mol (1997) in eastern Germany in deposits of similar age. Middle Pleniglacial deposits were not found in this studied part of the Maas valley, possibly because either the sediments were not recognized or the Middle Pleniglacial sediments were eroded at the onset of the Late Pleniglacial.

The transition towards the coldest part of the Weichselian, the Late Pleniglacial, is marked in the Vecht valley by a distinct change in fluvial style from a laterally inactive, moderately energetic river towards a highly energetic braided river (Fig. 7.1). Aggradation and reworking by braided rivers prevailed during the Late Pleniglacial. During the later part of this period increasing aeolian deposition occurred in both the Maas and Vecht river valleys. Both rivers became less energetic, though still braiding, and occupied smaller parts of the floodplains while the abandoned parts of the floodplains became covered by aeolian sands. At first these aeolian sands were reworked by shallow flowing water. Later on the aeolian sands adhered to periodically moist surfaces and finally aeolian deposition took place on a dry surface.

The sedimentary environment changed at the onset of the warmer and wetter Lateglacial. Both rivers became confined to much smaller floodplains, they incised significantly and changed from braiding towards meandering (Fig. 7.1). The change in fluvial style was gradual in the Maas valley where small channels from the Pleniglacial braided river system became abandoned as the larger ones started to incise and became increasingly curved. This process continued until finally one meandering channel remained active which migrated laterally. The reconstruction of the Vecht shows a maximum incision shortly before 11.3 ka and a similar change in fluvial pattern from braided to meandering. Both rivers reacted to the Younger Dryas cooling by incising first and deposition sediment in a braided river system afterwards. The Maas and Vecht both changed from braided to meandering rivers at the start of the Holocene, whereby incision occurred in the beginning of the Holocene warming and sedimentation afterwards.

Summarizing, erosional phases occurred around 40 ka, 27 ka, 13-12 ka, 11 ka and 10 ka (C-14 years), which were periods of significant climate changes.

7.3 SEDIMENT BUDGETS IN THE MAAS VALLEY DURING THE BØLLING, ALLERØD, YOUNGER DRYAS AND HOLOCENE

In chapter 4 amounts of reworked, deposited or eroded sediment were calculated by using terrace surfaces, thickness of terrace sediments and a GIS. The Lateglacial rivers were highly energetic as they reworked large amounts of sediment. This is explained by large stream powers which were related to the peaked discharge regimes. The highest energetic fluvial environments were the braided river in the

Younger Dryas and the transitional river between braiding and meandering in the Bølling period. They reworked three times as much sediment as the meandering river in the Allerød showing that the transitional phase between braiding and meandering was still very dynamic. The first net erosion on a regional scale did not take place until the start of the Allerød by a meandering river. A net incision occurred in the Younger Dryas, as aggradation was not large enough to compensate for the deep incision at the onset of this period. The depth of this incision was probably influenced by the confinement of the river to a narrow floodplain in the previous period. The Holocene Maas reworked much less sediment than the Lateglacial Maas. After an incisional phase at the start of the Holocene mostly deposition took place which resulted in a net aggradation. The meandering Holocene Maas was much less dynamic than the meandering Maas in the Allerød which is attributed to the diminished stream power as discharges became much more regular and lower in magnitude in the Holocene.

7.4 REFLECTION OF RIVER STYLE CHANGES IN THE SEDIMENT PETROGRAPHY

The diminished reworking of sediment in the Maas valley in the Bølling and Allerød periods is reflected in the heavy mineral composition by an increase in fresh supplied Maas minerals and a decrease of Rhine - Maas sediment from the subsoil. The increasing influx of Maas minerals is not observed in the Younger Dryas sediment as a result of the deep incision whereby older, Rhine-derived sediments were reworked. This trend is also observed in the gravel analysis, although less explicitly.

7.5 FACTORS INFLUENCING RIVER DYNAMICS

In general three external factors determine the behaviour of rivers: base level changes, tectonic movements and climate. Changes in fluvial processes like those described here for the Maas and Vecht can be triggered by a change in one or more of these factors. Base level changes were unimportant, since the sea-level was some 100 m lower than today (Jelgersma 1966) and both river valleys were therefore located far from the sea.

The occurrence of tectonic activity did also not account for the changes in sedimentological and morphological processes in the Maas valley, which was demonstrated in chapter 3. The formation of several distinct terrace levels in a short period like the Lateglacial cannot be attributed to the gradual long-term uplift of the hinterland. A detailed analysis of Late Pleniglacial and Lateglacial terrace slopes in the Maas valley by means of multiple regression analysis showed that breaks in terrace slopes occurred on two terraces. This was attributed to tectonic activity in the Venlo Graben but the effect on the river dynamics and morphology was small. Locally, channels were deeper incised, but the river pattern did not change in steeper sloping areas. In addition, the highly sinuous meandering Maas in the Allerød and the braided Maas in the Younger Dryas had similar palaeo-floodplain slopes (23.5 and 25.4 cm/km) but contrasting river styles. This indicates even more that river styles in the Maas valley were not determined by floodplain slopes. So, tectonically induced breaks in floodplain slopes did not force the River Maas to change its pattern or to incise significantly. However, the longterm uplift of the hinterland of the Maas does explain the tendency of the Maas to incise and form terraces. This is contrary to the Vecht valley where only one Lateglacial terrace is recognized.

By eliminating sea level (or base level) and tectonic activity as the driving mechanisms for the observed fluvial changes in the Maas and Vecht valleys, the factor climate change appears to be most important. The occurrence of comparable fluvial changes during periods with distinct climate changes throughout North-

Western and Central Europe confirms the assumption that rivers responded to climate changes by incision, deposition or change in fluvial style (see e.g. Mol 1997). The reconstruction of climate changes during the Weichselian becomes increasingly accurate. Greenland ice-cores revealed rapid changes in oxygen isotope ratios which are interpreted as high resolution temperature indicators (Dansgaard et al. 1993, Johnsen et al. 1995). Not all climate signals from the ice cores are recognized on the continent. However, vegetational reconstructions, based on macro-fossil and pollen-analysis, periglacial evidence and other proxies are combined for instance by Huijzer and Isarin (1997) and Huijzer and Vandenberghe (1998) to establish a detailed climate reconstruction as well (figure 7.2). These reconstructions are especially detailed for the Lateglacial period, but they are less detailed when organic sediment is absent like in the Late Pleniglacial.

7.6 THE NON-LINEAR FLUVIAL RESPONSES TO CLIMATE CHANGE OF THE MAAS AND VECHT

7.6.1 Fluvial response to cooling

A prominent cooling event occurred at the start of the Weichselian Late Pleniglacial. Ice masses reached their maximum extent and a change from discontinuous to continuous permafrost took place in The Netherlands (Huijzer and Vandenberghe 1998). Shorter cooling phases, that were reflected by fluvial changes, took place from 41-38 ka (the Hasselo Stadial after Ran and Van Huissteden 1990) and from 11-10 ka (the Younger Dryas). In both periods a transition from seasonally frozen ground to discontinuous permafrost occurred (Huijzer and Vandenberghe 1998, Isarin 1997).

The erosional phase and fluvial style change from a lower energetic, most probably anastomosing river before circa 41 ka into a higher energetic river after 36 ka in the Vecht can most likely be explained by the cooling in the Hasselo Stadial. It seems as if thresholds were crossed in the cold Hasselo Stadial and the river adapted a higher energetic fluvial style. The Hasselo Stadial is the only stadial in this period where a change from seasonal frost to discontinuous permafrost is recognized and it is also the only Stadial in which a significant effect on vegetation was observed (Ran 1990). The period prior and following the Hasselo Stadial is characterized in the isotope curves as an unstable period in which several short cold and warm spikes are present. The overall tendency during the Middle Pleniglacial (despite of warm and cold spikes) was a decreasing temperature (Fig. 7.2). As a result the river did not return to its previous system.

The Middle Pleniglacial river plains were occupied by an arctic shrub to tundra vegetation (Ran and Van Huissteden 1990). In present discontinuous permafrost areas evapotranspiration rates are lower than in areas of seasonal frost as more heat is needed to melt ice and consequently less heat is left for evapotranspiration (Woo and Winter1993). Using this as an analogue it could be stated that during the Hasselo Stadial evaporation decreased. A lower evapotranspiration resulted in an increased discharge versus sediment supply which led to erosion and increased floods at the start of the Hasselo Stadial (circa 40 ka). Another feature that might have occurred is an increasing thickness of the river ice which resulted in a higher energetic fluvial regime during the ice breakup in spring. The increased floods in turn favoured deposition of sand in the floodplain which, together with a water level rise could be the cause of the ending of marsh vegetation growth in the floodplain (Ran 1990). When the vegetation was sufficiently destroyed more sediment could be supplied to the river and the observed change in fluvial style, from a lower to a higher energetic bit laterally inactive river type took place after 40 ka. Besides, the interfluves were probably not densely vegetated, and formed a sediment source to the river system, which may be derived from the occurrence of aeolian sands in the Dinkel valley during the Middle Pleniglacial (Van Huissteden 1990).

The cold Late Pleniglacial is characterized by the presence of braided rivers in most European lowland rivers (Mol, 1997). In the Vecht valley a change took place from a moderately-energetic, but laterally inactive river in the Middle Pleniglacial towards a highly energetic braided river in the Late Pleniglacial. The severe cold conditions resulted in a barren landscape, where vegetation must have been very limited or absent as generally no organic sediments are found from this period. Protection against erosion was minimal so sediment supply was large. This, together with peaked discharges resulted in a braided river. The change towards a braided river occurred some time after 31 ka in the Vecht valley.

The onset of the Younger Dryas period is characterised by a rapid decrease in temperature (Fig. 7.2) and wetter conditions as indicated by higher lake levels (Bohncke et al. 1987). The combination of decreased evaporation rates, decreased soil water storage capacity and the establishment of discontinuous permafrost resulted in increased peak discharges. These fast changes in hydrology occurred when vegetation was intact at first and soil and bank stability were maintained, which led to erosion. As soon as the (forest) vegetation was destroyed and changed to a more open shrub with heather vegetation (Bohncke et al. 1993), the sediment load increased and a braided river pattern developed. The depth of incision in the Maas valley was larger than in the Vecht valley and is possibly related to the larger catchment area of the Maas or to the stronger uplift of the hinterland. The fluvial response of other European rivers to the Younger Dryas climate change is diverse. A similar incision and change towards higher energetic conditions is recognized in lowland Britain (Rose 1995), in the Ems and Niederrhein valleys (Klostermann 1995) and in parts of the upper Vistula (Kalicki and Zernickaya 1995). Other rivers did not change their fluvial style like the Scheldt (Kiden 1991), the Warta (Vandenberghe et al. 1994) and large parts of the Vistula (Starkel 1995). Apparently the cooling event was not severe or long enough to cause incision and changes in fluvial styles in all river valleys.

7.6.2 Fluvial response to warming

Two major warming phases took place: at the start of the Lateglacial and the Holocene (Fig. 7.2). The Lateglacial was probably also wetter as indicated by high lake levels (Bohncke and Wijmstra 1988). Both the Maas and the Vecht reacted by incision and by a change in fluvial style towards meandering. The fluvial reconstructions are most complete and detailed from the Maas valley, where a gradual change is seen from a braided river in the Late Pleniglacial and early Bølling (chapter 4) via a transitional phase between braided and meandering and finally towards a meandering river at the start of the Allerød. In the Vecht valley deposits from a transitional phase have not been detected. Only deposits from a meandering river have been found that were dated at 11.320 C14 yr BP, which indicates that incision took place at least before the middle Allerød. Similar changes in fluvial style are found in many other European river valleys (Kiden 1991, Collins et al. 1996, Klostermann 1995, Kozarski 1983, Szumanski 1983). At a few occasions a transitional phase between braiding and meandering has been distinguished (Lefevre et al. 1995, Vandenberghe et al. 1994, Schirmer 1983, Lipps and Caspers 1990).

This fluvial change can be explained by the increasingly better protection of the soil against erosion by the Lateglacial vegetation development and by regularization of the seasonal run off pattern, due to larger soil water storage capacity. A herbaceous vegetation with dwarf bushes colonised the surface in the early Bølling, which resulted in a larger soil surface stability than in the previous period and sediment load diminished. The changed sediment to discharge ratios resulted in the tendency of the large channels to incise, while smaller ones became abandoned. During the Bølling, the vegetation became increasingly denser. In the middle to late Bølling birch colonised the surface (Paus 1995, Hoek 1997a) and open birch

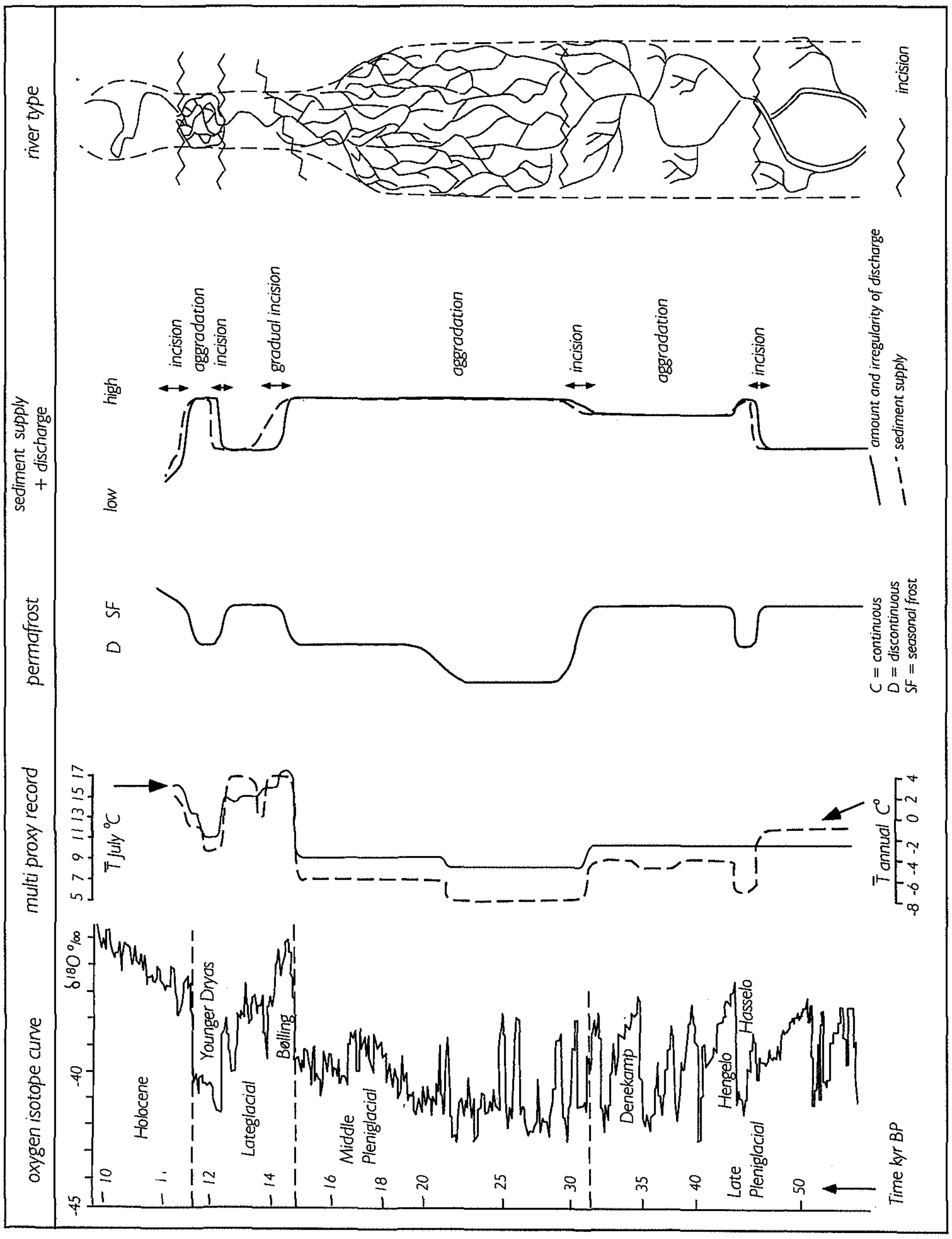


Figure 7.2 Compilation of river style changes, oxygen isotope curve mean July temperature based on multi proxy records, permafrost and sediment and discharge supply. Oxygen isotope curve after Dansgaard et al. (1993); mean July temperature for Lateglacial after Hoek (1997a); mean July temperature for Pleniglacial after Huijzer and Vandenberghe (1998); permafrost after Huijzer and Vandenberghe (1998) and Isarin (1997).

woods developed. The change from an almost bare landscape to a vegetation of herbs and shrubs towards a forest vegetation influenced erosion rates greatly (chapter 2) and resulted in a decreasing sediment supply.

The fluvial response to the Lateglacial warming is thus a gradual change in fluvial style and incision. The maximum depth of incision was not at the onset of the warming phase, but after a time lag in which vegetation needed time to develop and protect the soil. This time-lag was large at the onset of the Lateglacial because vegetation was nearly absent in the previous period. The Holocene warming is characterized by a much faster restoration of the vegetation which resulted most probably in a faster change in fluvial style. Evidence for a transitional phase between the braided river in the Younger Dryas and the meandering one in the Holocene was not found in both river valleys. The discharge characteristics in the Holocene were also different from those in the Bølling period. The mean annual temperatures rose quickly at the start of the Holocene, just as at the start of the Bølling. The winter temperatures remained low, however, in the Bølling (Bohncke and Vandenberghe 1991) so that snowfall in winter was high and the soil was frozen for major parts of the year. The discharges in the Bølling were consequently irregular and depended for a large part on snowmelt, while in the early Holocene, discharge was determined less by snowmelt and more regularly distributed throughout the year. In addition, soil water storage in the Bølling was relatively low when compared to the early Holocene. The more regular discharge, combined with the low slope erosion rates which is linked to the rapid vegetation development, favoured the accelerated change to a meandering river in the Holocene.

7.6.3 Fluvial response to increased aridity

During the Late Pleniglacial an increased influence of aeolian processes in the river flood plains is observed. Both the Vecht and Maas occupied an increasingly narrower floodplain and the abandoned parts of the floodplains became covered by aeolian sands. The transition from fluvial to aeolian deposition is gradual as is demonstrated in the Vecht valley in the Nolderveld exposure (chapter 6). The braided river system in the Maas valley became less dynamic as smaller scale channel fills and finer grained sediments are found in comparison to the previous period (chapter 2). The increase of windblown sands into the river flood plains is related to the severe aridity that followed the maximum extent of the ice-sheet around 20 ka. The vegetation cover and protection against erosion was minimal during this period enhancing wind erosion. The increased aridity decreased the water supply to the rivers and smaller parts of the floodplains were in use by less dynamic braided rivers. The fluvial style remained braided, but the channel dimensions became smaller (at least in the Maas valley) and the braided river became less dynamic.

7.7 MODEL FOR LOWLAND RIVER RESPONSES TO CLIMATE CHANGES

The assumption that rivers incise during warm periods and aggrade during cold periods is too simple. Various, increasingly complex models have been developed to explain fluvial behaviour in response to climate change or other factors (see chapter 1). Vandenberghe (1993) explained the complex fluvial responses to climate changes in a non-linear model. Incisions took place in rather short time intervals when climate changes took place, from cold to warm and vice versa. At the start of a cold period evaporation decreased which led to increased discharges while vegetation remained intact at first so that bank and soil stability was maintained. The increased water versus sediment ratio caused erosion. When vegetation became destroyed and sediment supply increased aggradation occurred. At the onset of a warm period vegetation quickly stabilized the soil and reduced sediment supply, which resulted also in incision. These processes were thought to occur when internal morphological thresholds of rivers were exceeded.

The non-linearity of fluvial processes in response to climate change is not contradicted in this study. The short phases of instability that occur during climate changes prove to be very important as significant erosion and changes in fluvial style took place. Stable phases are phases in which either no significant climate changes occurred, or climate changes occurred, but internal thresholds of the river were not crossed. Some climate changes were so distinct (such as the onset of the Late Pleniglacial, Lateglacial and Holocene) that fluvial responses are recorded from almost all European lowland river valleys (Mol 1997). Other climate changes were too short or not explicit enough (onset Hasselo Stadial and Younger Dryas) so that internal thresholds were crossed by some, but not all rivers. When internal thresholds are crossed, the rivers react in a similar manner to the climate changes. The non-linear model of Vandenberghe (1993) describes these reactions generally correct, but a refinement to the model may be considered.

Firstly, the non-linear model of Vandenberghe (1993) is refined by a different explanation for incision at the start of a warm period. The incision begins as soon as vegetation starts to protect the surface and progrades as vegetation develops in a denser cover, preferably with trees which establish bank and soil stability. The maximum depth of incision is reached as the vegetation develops into a dense, woody cover. The time vegetation needed to develop into a dense tree cover varied. At the onset of the Lateglacial it took some 1000 years, while the vegetation development was much faster at the onset of the Holocene. Combined with a more regular discharge regime in the Holocene, and larger soil water storage capacity it explains why a transitional phase between braiding and meandering is seen in the Lateglacial but not in the Holocene. The role of the vegetation development is thus more important than envisaged by Vandenberghe (1993).

Secondly, this study showed that fluvial response to warming can be more gradual than a fluvial response to cooling. The incision as a response to warming can be gradual and progressive, while the incision at the onset of a cold period is abrupt and relatively large, as is seen in the responses of the Vecht and Maas to the Younger Dryas or Hasselo Stadials. The different responses can be explained by the vegetation development. The time necessary for vegetation to develop at the start of a warm period varies, but may take a long time on a barren surface such as at the onset of the Lateglacial where the maximum depth in the Maas valley was reached some 1000 years after the warming event itself. The fluvial response to cooling is always fast which is explained by the sudden changes in discharge characteristics, resulting in changed sediment-water ratios and incision; followed by a quickly destroyed vegetation cover. The destruction of a vegetation cover does not take much time, leads to fast changes in sediment-water ratios and consequently a rapid change in fluvial style.

7.8 DISCUSSION ON THE NATURE OF INTERNAL FLUVIAL THRESHOLDS

The rivers responded only to climate changes when internal thresholds were crossed. An important factor determining changes in fluvial processes is the vegetation cover in the floodplain and on the interfluves. Climate changes that had no significant effect on the vegetation are not reflected by fluvial change; whereas the climate changes that affected the vegetation cover resulted in fluvial responses. Oxygen isotope curves from the Greenland ice-sheet show multiple cold and warm periods in the Middle Pleniglacial (Fig. 7.2) which were not followed by a change in vegetation cover (Ran 1990) and in fluvial response (Kasse et al. 1995b). The cold period in the Middle Pleniglacial that had a significant impact on the vegetation cover was the Hasselo Stadial, which resulted in a destruction of the vegetation cover in the river plains and probably also on the interfluves and is reflected in the development of the Vecht by an incisional phase and a change in fluvial style.

A comparison of the Vecht and Dinkel (an upstream tributary of the Vecht) showed that only the Vecht changed its fluvial style as a response to the cooler Hasselo Stadial. The erosional phase around 40 ka is recognized in both rivers but a change towards higher energetic fluvial conditions has not been found in the Dinkel valley (Van Huissteden 1990). Apparently the internal thresholds were crossed in the Vecht valley, but not in the upstream Dinkel valley. A hypothetical explanation is that these thresholds are related to catchment size. The larger catchment of the Vecht induced a stronger impact of discharge changes and a more widespread destruction of vegetation in the river floodplain. The wider floodplain served also as a larger depositional area for aeolian sands and thus for a larger supply of sediment. Evidence for a low energetic Middle Pleniglacial Maas has never been found, either because sediments from this river were eroded during the Late Pleniglacial or because the Maas remained braided. Mol (1997) summarized multiple river styles in European lowland valleys and reported low-energetic or ephemeral anastomosing rivers but also braided rivers during the Middle Pleniglacial.

The factors valley slope and grainsize are thought to have been unimportant in determining the crossing of internal thresholds, which will be illustrated by examples from the Maas and Vecht. The crossing of thresholds by the Maas in the Lateglacial was not determined by changes in valley slopes. Contrasting river styles, such as the highly sinuous meandering river in the Allerød and the braided river in the Younger Dryas had similar valley slopes. Furthermore, local steepening of the valley slope such as on the Bølling terrace, where slopes increased from 28 to 40 cm/km did not result in a change of fluvial style. Grainsize was most probably neither an important factor in determining whether thresholds were crossed or not. Firstly, the Vecht changed its pattern after circa 40 ka and the Dinkel did not although the sediment grainsizes in both valleys are very similar. Secondly, the Maas and Vecht showed comparable developments from the Late Pleniglacial until the Holocene although the average grainsize in the Maas valley is considerably coarser.

The importance of the vegetation cover on fluvial development was great and its development or destruction determined for a great part whether a river crossed internal geomorphological thresholds leading to incision or aggradation or changes in style. It is therefore important to have detailed, well-dated vegetation reconstructions. The large Weichselian climate changes affected the vegetation cover everywhere which resulted in a similar reaction of and nearly all European rivers in a similar manner. Smaller climate changes did not affect the vegetation cover everywhere in a comparable way which explains the local response of some rivers only.

7.9 CONCLUSIONS

The Maas and Vecht reacted in a non-linear manner to climate changes. Incisions took place at climate changes from cold to warm and vice versa. The fluvial response to cooling is abrupt and incision is explained by the delayed destruction of the vegetation cover as opposed to fast changes in discharge. As soon as the vegetation becomes destroyed and sediment is supplied to the river the incision stops and aggradation follows. The fluvial response to warming is more gradual. Incision occurs when sediment supply becomes restricted by the development of a vegetation cover. The rate of vegetation development determines the changes in fluvial style.

The role of the vegetation cover was very important as it determined the availability of sediment into the river but also soil water storage capacity and evapotranspiration rates and therefore discharge magnitudes. Changed sediment-water ratios determined the observed fluvial style changes and incisional phases in the Maas and Vecht valleys. Not all Weichselian climate changes were reflected by changes in vegetation. Only the climate changes that affected the vegetation resulted in a fluvial response.

The importance of tectonic movements in triggering fluvial changes proved to be insignificant as the Maas did not change its fluvial style in response to tectonically induced changes in valley slope and similar valley slopes were found in both meandering and braided river systems.

The calculation of the amounts of reworked sediment and the net amount of incision or deposition during three Lateglacial periods and the Holocene confirmed the fluvial reconstructions made so far. The transitional phase of the Maas between braiding and meandering proved to be very dynamic and large amounts of sediment were reworked but net erosion did not occur until the end of the Bølling period. This stresses again the importance of the vegetation development as net erosion only occurred after the vegetation had evolved into a dense, woody cover. The deep incision of the Younger Dryas river, is explained by the confinement of the river in a small floodplain in the previous Allerød period. The deep incision at the start of the Younger Dryas was reflected in the sediment petrography.

Summary

INTRODUCTION

The effect of past climate changes on two Dutch rivers is the subject of this thesis. The last four periods of well expressed climatic change are of particular interest and occurred around 27, 13, 11 and 10 C14 ka BP. They mark the transition from the Weichselian Middle Pleniglacial to the phase of maximum cold in the Late Pleniglacial, the transition to the warm Lateglacial and the onset and ending of the last cold stadial, the Younger Dryas. Climate and vegetation reconstructions of these periods are well documented. Distinct changes in fluvial processes took place at those times. By studying two Dutch rivers an attempt is made to describe and explain fluvial changes in relation to local factors (like tectonic setting) and regional factors (climate changes).

The study areas are located in the Maas and Overijsselsche Vecht valleys, where Pleniglacial and Lateglacial terraces are found. Most sedimentological information is obtained from boreholes and a limited amount of pits. Detailed analysis of 1:10,000 topographical maps provide morphological information.

LATEGLACIAL SEDIMENTOLOGICAL AMD MORPHOLOGICAL CHANGES IN A LOWLAND RIVER IN RESPONSE TO CLIMATIC CHANGE; THE MAAS, SOUTHERN NETHERLANDS

The Maas repeatedly incised in an increasingly narrower floodplain which resulted in the formation of one Pleniglacial and three Lateglacial terraces. Channel scars on top of these terraces, combined with sedimentological information, show distinct changes in fluvial styles. The Maas changed from a highly energetic, aggrading braided river to a slightly less energetic, but still braided river in the Late Pleniglacial. Parts of the floodplain were abandoned and became covered by aeolian sands. During the Bølling period the braided river changed into a system transitional between braiding and meandering. The larger channels incised and sinuosity increased, while minor channels were abandoned. This process continued until one, meandering channel remained active in the Allerød. The Maas incised deeply and changed into a braided river in the Younger Dryas. Another incision took place at the onset of the Holocene, accompanied with a change in style towards meandering.

The fluvial style changes and phases of incision and deposition may most likely be explained by the distinct climate changes that took place. This is supported by a comparison with other European lowland rivers which showed similar fluvial changes in the same periods. Erosion occurred at the transition from a warm to cold period and vice versa, while aggradation occurred during cold or warm periods conform the non-linear model of Vandenberghe (1993, 1995). Incision at the onset of a cold period is explained by the immediate change in discharge regime in combination with a delayed change in sediment supply. Discharges became larger and more irregular, while the presence of a vegetation cover restricted sediment supply. Incision as a response to warming is explained by the diminishing sediment load related to increasing vegetation cover and more regular discharges (influenced by an increasing in soil water storage capacity).

The delayed response of the Maas to climate change around 12-13 ka was thought to be an intrinsic evolution over some 1300 years (Vandenberghe 1995). This gradual change in pattern can, however, also be explained by a direct response of the river to changing vegetation cover, discharge characteristics and sediment load. The sudden changes in fluvial style at the Allerød - Younger Dryas and Younger Dryas - Holocene transitions can be explained by the rapid changes in vegetation cover and discharge characteristics, while the gradual change at the start of the Lateglacial is related to gradual changes in sediment supply, combined with a gradually changed discharge regime.

TECTONIC VERSUS CLIMATIC CONTROLS ON THE RIVER MAAS DYNAMICS DURING THE LATEGLACIAL

A better understanding of the tectonic control on the fluvial changes of the Maas is achieved by constructing longitudinal profiles of each Maas terrace using multiple linear regression analysis. This provides an objective way of determining palaeo valley slopes. The reliability of the calculations proved to be high as the R squared varied from 0.92 to 0.99. The terraces from the late Pleniglacial and Bølling periods are relatively steep (41.8 cm / km and 35.1 cm / km) when compared to the other Lateglacial terraces (23.5 and 25.1 cm / km). The Allerød and Younger Dryas terraces can be correlated nicely with observed values downstream. However, a break in slopes occurs when the Late Pleniglacial and Bølling terraces are compared with areas downstream. Both terraces were studied in more detail to determine the location of the break in slope. Each terrace was divided into four parts and a multiple linear regression analysis was done for each part of the terrace. It appeared that the slope of the Bølling terrace decreased drastically from 34.8, 36.2 and 39.6, cm / km in 27 cm / km in the most downstream part of the terrace. This decrease in slope was also found fot the late Pleniglacial terrace.

The steeper slope segments and breaks in slopes of the two oldest terraces are probably best explained by tectonic movements in this area. The influence on the fluvial style was small, however, since no change in fluvial morphology was observed into areas of steeper valley slopes. Furthermore, the slopes of the Allerød and Younger Dryas terraces are quite similar while the fluvial styles are quite different: a highly sinuous meandering river versus a braided one. This illustrates that the influence of tectonic movements, inferred from longitudinal terrace profile analyses, was probably restricted on the river evolution of the Maas during the Lateglacial. The phases of incision and deposition are therefore better explained by the distinct climate changes that affected sediment and water supply significantly.

LATEGLACIAL RIVER SEDIMENT BUDGETS IN THE MAAS VALLEY; THE NETHERLANDS

In this chapter the amounts of eroded, deposited and reworked sediment during each Lateglacial period were calculated in the Maas valley. The sediment budgets allow a comparison of the transport capacity of the different river styles and help to explain the observed fluvial changes. Terrace extensions and thickness of sediment bodies were combined in a GIS, to enable a three-dimensional approach. The study area was divided into grid cells of 200 by 200 meter. Minimum and maximum possible terrace extensions were calculated by adding or subtracting parts of the valley to the palaeo-floodplains. The thickness of sediment bodies was obtained by using extra mechanical borehole logs that penetrated the base of the sediment bodies.

The amounts of reworked sediment were comparable for the Allerød and Holocene periods, while three times as much sediment was reworked in the Bølling and Younger Dryas periods. This might be attributed to the higher energetic rivers in the Bølling (transitional system between braiding and meandering) and Younger Dryas (braiding) versus the meandering rivers of the Allerød and Holocene. Since the Allerød period was some ten times shorterin length than the Holocene the meandering Maas in the Allerød was much more dynamic than the Holocene Maas. Subtraction of palaeo-floodplain altitudes provide information on processes of sedimentation or incision between periods. The first net incision on a floodplain scale took place by the meandering river in the Allerød, after the vegetation development reduced sediment load significantly. A deep incision at the onset of the Younger Dryas was not compensated by sedimentation afterwards which resulted also in a net incision. The amount of reworked sediment in the Younger Dryas was similar to the Bølling period, suggesting that stream powers were comparable. However, the river was confined to a narrow floodplain in the Allerød, and bank stability was high due to the denser vegetation cover. This might explain why the excess stream power was spent on bed erosion rather than lateral erosion.

HEAVY MINERAL AND GRAVEL COMPOSITION OF THE LATEGLACIAL MAAS SEDIMENTS

Gravel and heavy mineral analyses were performed to determine whether the fluvial style changes of the Maas, with the associated phases of incision and deposition are reflected in the sediment petrography. The substratum in the Maas valley consists of a mixture of sediment derived from both Rhine and Maas. During the Pleniglacial, Lateglacial and Holocene this mixed Rhine-Maas sediment was enriched by pure Maas sediments. During the Pleniglacial and Bølling the Niers, a tributary of the Maas, served as a Rhine-branch supplying Rhine sediments to the Maas valley. The sediment in the Maas valley is thus a complex mixture from Maas and Rhine derived sediments and requires very detailed analyses.

During the Lateglacial an increasing supply of Maas minerals is observed with decreasing age in the successive terraces, both up- and downstream of the Niers confluence. This is shown by an increase in tourmaline, stable- and opaque minerals and a decrease in garnet and alterite. It illustrates the diminished amount of reworking of older, mixed Rhine-Maas sediment during the Lateglacial. This trend is also confirmed by the gravel analyses. The increasing influx of Maas sediments is not observed in the Younger Dryas deposits as a result of the deep incision during this period whereby older, Rhine-derived sediments were reworked. The source area of the Younger Dryas dunes on the east banks of the Maas was the Younger Dryas floodplain as the heavy mineral composition of the Younger Dryas terrace matches that of the Younger Dryas dunes, both up- and downstream of the Niers-Maas confluence.

CHANGES IN RIVER STYLES IN RESPONSE TO WEICHSELIAN CLIMATE CHANGES IN THE EASTERN NETHERLANDS

In the Vecht valley, eastern Netherlands, distinct changes in fluvial style and erosional phases took place from the Pleniglacial to the Holocene. The river changed from a low energetic, anastomosing river in the Weichselian Middle Pleniglacial to a braided river in the Late Pleniglacial. Furthermore, a change from fluvial to aeolian-dominated sedimentation occurred during the Late Pleniglacial. In the Lateglacial the river incised and became a low-sinuosity meandering river at first and later on a braided river in the Younger Dryas period. These fluvial changes are accompanied by erosional phases and both are most likely the result of climatically induced changes in the water - sediment discharge ratio. Vegetation appears to be an important factor as it determined bank and soil stability and therefore sediment supply, but also evapotranspiration rates, which influenced discharges characteristics. A model explaining the fluvial response to climate change is presented and discussed by using the Vecht river as an example.

Samenvatting

INTRODUCTIE

In dit onderzoek wordt het effect van klimaatsveranderingen op twee rivieren in Nederland bestudeerd. De laatste vier grote klimaatsveranderingen markeren de overgang van het Weichsel Midden Pleniglaciaal naar de maximale koude in het Laat Pleniglaciaal (27), de overgang naar het warme Laatglaciaal (13) en het begin (11) en einde (10 duizend ^{14}C jaar) van de laatste koude periode: de Jonge Dryas. Van deze perioden zijn goed gedocumenteerde vegetatie- en klimaatsreconstructies beschikbaar. Specifieke veranderingen in fluviatiele processen vonden plaats en door deze van twee rivieren te bestuderen kan een onderscheid gemaakt worden tussen lokale factoren (zoals tektonische geschiedenis) of regionale (klimaatsveranderingen) factoren op de rivierevolutie. De studiegebieden bevinden zich in delen van de Maas en Overijsselsche Vecht vallei, waar pleniglaciale en laatglaciale terrassen zijn gevonden. Sedimentologische informatie kwam voornamelijk uit boringen en een beperkt aantal ontsluitingen, terwijl morfologische informatie werd verkregen uit gedetailleerde analyses van 1:10.000 topografische kaarten.

LAATGLACIALE SEDIMENTOLOGISCHE EN MORFOLOGISCHE VERANDERINGEN IN EEN LAAGLAND RIVIER ALS REACTIE OP KLIMAATSVERANDERINGEN; DE MAAS, ZUID NEDERLAND

De Maas heeft zich herhaaldelijk ingesneden in een steeds smaller wordende vallei wat geresulteerd heeft in de vorming van één pleniglaciaal en drie laatglaciale terrassen. Veranderingen in rivierpatroon zijn af te leiden uit de combinatie van restgeulen op deze terrassen en sedimentologie. De Maas veranderde van een hoog-energetisch, aggraderend vlechtend systeem in een iets minder energetisch maar nog steeds vlechtend systeem in het Laat Pleniglaciaal. Delen van de riviervallei raakten buiten gebruik en bedekt met eolische zanden. In de Bølling periode veranderde het vlechtende systeem in een overgangssysteem tussen vlechtend en meanderend. De grootste geulen sneden zich in en werden hoger sinuoos terwijl de kleinere verlaten werden. Dit proces leidde uiteindelijk tot één actieve, meanderende geul in de Allerød. De Maas sneed zich in de Jonge Dryas diep in en veranderde opnieuw in een vlechtende rivier. Een volgende insnijding en omslag naar een meanderend patroon vond plaats aan het begin van het Holoceen.

De veranderingen in rivierpatroon, insnijding en accumulatie kunnen waarschijnlijk het beste verklaard worden door de grote klimaatswisselingen. Een vergelijking met andere Europese laaglandrivieren laat vergelijkbare fluviatiele veranderingen in dezelfde perioden zien, wat deze veronderstelling ondersteunt. Erosie vond plaats op de overgang van een warme naar een koude periode en vice versa, terwijl aggradatie optrad gedurende warme of koude periodes, conform het niet-lineaire model van Vandenberghe (1993, 1995). Insnijding aan het begin van een koude periode kan verklaard worden doordat afvoeren groter en onregelmatiger werden terwijl sediment aanvoer beperkt bleef door de aanwezige vegetatie. Insnijding als gevolg van een opwarming ontstond door een afnemende sediment aanvoer, gerelateerd aan een dikker wordend vegetatiedek en regelmatiger afvoeren (beïnvloed door een steeds groter wordende bergingscapaciteit van de bodem). De vertraagde respons van de Maas rond 12 - 13 kj werd gezien als een intrinsieke evolutie die zo'n 1300 jaar duurde (Vandenberghe 1995). De graduele verandering in rivierpatroon kan echter ook worden verklaard door een direkte reactie van de rivier op veranderende afvoerkarakteristieken en sedimentlast. De plotselinge veranderingen in rivierpatroon op de overgang tussen Allerød - Jonge Dryas en Jonge Dryas - Holoceen zijn te verklaren door de snelle veranderingen in sedimentlast en afvoerregime.

DE INVLOED VAN TEKTONISCHE VERSUS KLIMATOLOGISCHE FACTOREN OP DE LAATGLACIALE MAAS DYNAMIEK

De invloed van tektonische activiteit op de Maas is gereconstrueerd door longitudinale gradiëntlijnen (verhanglijnen) te berekenen met behulp van meervoudige lineaire regressie analyse. Dat maakt het mogelijk om op een objectieve manier de paleo-verhanglijnen te bepalen. De betrouwbaarheid van de analyses bleek goed te zijn met R^2 waarden tussen 0.92 en 0.99. De pleniglaciale en bølling terrassen zijn relatief steil (41,8 en 35,1 cm/km), vergeleken met de andere Laatglaciale terrassen (23,5 en 25,1 cm/km). De Allerød en Jonge Dryas terrassen konden goed gecorreleerd worden met waarden stroomafwaarts, in tegenstelling tot de Laat Pleniglaciale en Bølling terrassen die elk een knik in de verhanglijnen lieten zien. Beide laatste terrassen zijn daarom in meer detail bestudeerd om erachter te komen waar precies de verhanglijnen knikken. Elk terras werd verdeeld in vier delen waar afzonderlijke regressie analyses werden uitgevoerd. Het verhang van het Bølling terras neemt drastich toe van 27 cm/km in het meest stroomafwaartse deelgebied naar 39,6, 36,2 en 34,8 cm/km in de andere deelgebieden. Deze toename van het verhang is ook te vinden bij het Pleniglaciale terras.

De steilere verhanglijnen en de knikken in de verhanglijnen van de twee oudste terrassen kunnen waarschijnlijk het beste door tektonische activiteit in dit gebied verklaard worden. De invloed van tektoniek op het rivierpatroon was klein, aangezien er a. geen verandering in riviermorfologie te zien is tussen gebieden met een verschillend verhang en b. de verhanglijnen van de Allerød en Jonge Dryas terrassen vrijwel gelijk zijn terwijl de rivierpatronen nogal verschillend zijn: een hoog sinuose meanderende rivier tegenover een vlechtende rivier. Dit illustreert de geringe invloed van tektoniek op de Laatglaciale Maas. De insnijdings- en afzettingsfases kunnen beter worden verklaard door klimaatsveranderingen die in belangrijke mate de aanvoer van water en sediment bepaalden.

LAATGLACIALE RIVIER SEDIMENT BUDGETTERINGEN IN DE MAAS VALLEI, NEDERLAND

In dit hoofdstuk zijn de hoeveelheden geërodeerd, afgezet en omgewerkt sediment tijdens elke Laatglaciale periode in de Maas vallei berekend. De sedimentbudgetten maken het mogelijk om een vergelijking te maken van de transportcapaciteiten van de verschillende rivierpatronen. Dit kan helpen om de verschillende fluviatiele veranderingen te verklaren. Terrasoppervlakken zijn gecombineerd met de dikte van sedimentpakketten in een GIS (Geografisch Informatie Systeem), volgens een drie-dimensionale aanpak. Het studiegebied is verdeeld in gridcellen van 200 bij 200 meter. De terrasoppervlakken zijn berekend door paleorivierdalen van elkaar af te trekken of op te tellen. De dikte van sediment pakketten is bepaald met behulp van mechanische diepboringen.

Er werden vergelijkbare hoeveelheden sediment omgewerkt in het Allerød en Holoceen, terwijl er drie keer zoveel werd omgewerkt in de Bølling en Jonge Dryas. Dit komt waarschijnlijk door de hoger energetische Maas in de Bølling (overgang tussen vlechtend en meanderend) en Jonge Dryas (vlechtend) versus en meanderende Maas in de Allerød en Holoceen. De meanderende Allerød Maas blijkt dus veel dynamischer dan de Holocene Maas geweest te zijn aangezien de Allerød periode zo'n tien keer korter duurde. Wanneer de hoogte van de riviervalleien van elkaar worden afgetrokken kan de insnijding of aggradatie tussen twee perioden berekend worden. De eerste insnijding op grote schaal vond plaats in het Allerød, door de meanderende rivier, nadat de uitbreidende vegetatie de sediment toevoer beduidend had verminderd. De diepe insnijding aan het begin van de Jonge Dryas werd niet gecompenseerd door sedimentatie naderhand wat resulteerde in een netto insnijding. De hoeveelheid omgewerkt sediment in de Jonge Dryas lijkt op die in de Bølling periode, wat suggereert dat de transportcapaciteit van beide rivieren vergelijkbaar was. De rivier was echter beperkt tot een smalle riviervlakte in de Allerød, en de oeverstabiliteit was hoog door het dichte vegetatiedek. Dit verklaart waarschijnlijk waarom de extra transportcapaciteit werd aangewend om vertikaal te eroderen in plaats van lateraal.

ZWARE MINERALEN EN GRIND INHOUD VAN LAATGLACIALE MAAS SEDIMENTEN

Om te zien of de veranderingen in rivierpatroon en de daarmee samenhangende fases van insnijding en aggradatie terug te vinden zijn in de sedimentpetrografie zijn er grind en zware mineralen analyses uitgevoerd. Het substraat in de Maas vallei bestaat uit een mix van Rijn en Maas sediment. Tijdens het Pleniglaciaal, Laatglaciaal en Holoceen is deze mix verrijkt met pure Maas sedimenten. Voorts zijn er door de Niers, als voormalig deel van het Rijn-systeem, Rijn sedimenten aangevoerd tijdens het Pleniglaciaal en Bølling. De afzettingen in de Maas vallei zijn dus complex en vragen een zeer gedetailleerde analyse.

Tijdens het Laatglaciaal is een toenemende aanvoer van Maas sedimenten te zien gaande van de oude naar jongere terassen, zowel stroomop- als stroomafwaarts van de Niers. Dit is te zien in de toename van toermalijn, stabiele en opake mineralen en de afname van granaat én alteriet. Dit illustreert de afnemende omwerking van ouder, gemixed Maas-Rijn sediment. Deze trend wordt ondersteund door de grindanalyses. De toenemende invloed van Maas sediment treed niet op in de Jonge Dryas afzettingen, omdat een diepe insnijding aan het begin van de Jonge Dryas ervoor zorgde dat ouder Rijn sediment omgewerkt werd. Het brongebied voor de Jonge Dryas duinen aan de oostzijde van de Maas is de Jonge Dryas riviervlakte geweest, omdat de zware mineralen samenstelling van de duin- en rivierzanden op elkaar lijkt, zowel stroomop- als stroomafwaarts van de Niers.

VERANDERINGEN IN RIVIERPATROON ALS GEVOLG VAN KLIMAATSVERANDERINGEN IN HET WEICHSEL IN OOST NEDERLAND

Specifieke veranderingen in rivierpatroon en erosie fases hebben in de Vecht vallei plaatsgevonden van het Midden Pleniglaciaal tot in het Holoceen. De rivier veranderende van een laag-energetische, lateraal niet migrerende rivier in het Midden Pleniglaciaal in een vlechtende rivier in het Laat Pleniglaciaal. Verder vond er een omslag plaats van rivier- naar eolisch gedomineerde sedimentatie tijdens het Laat Pleniglaciaal. In het Laatglaciaal sneed de Vecht zich in en werd eerst een laag-sinuose meanderende rivier en later, in de Jonge Dryas. Deze fluviatiele veranderingen hangen samen met insnijdingsfases en zijn zeer waarschijnlijk het resultaat van klimaatsgestuurde veranderingen in de water / sediment ratio. De vegetatie lijkt een belangrijke rol te spelen, omdat deze oever- en bodemstabiliteit beinvloedt en daardoor de beschikbaarheid van sediment. De vegetatie bepaalt echter ook verdamping en bergingscapaciteit van de bodem en daarmee de afvoerkarakteristieken. Een model dat de fluviatiele veranderingen als gevolg van klimaatsveranderingen verklaart wordt besproken en uiteen gezet met de Vecht als voorbeeld.

Appendix Petrographical analyses in the Maas and Niers valleys

Table 1 Gravel composition of Weichselian and Holocene deposits in the Maas and Niers valleys (this study), in percentages. MQ= milky quartz, RTQ= red transparent quartz, TQ= transparent quartz, FL= flint, P= porfyry, FS= feldspar, QF= quartz and feldspar, C= crystalline, LY= lydite, D= diverse, GS= green sandstone, LS= limestone, S= total sandstone.

sample		X-coord	Y-coord	n	MQ	RTQ	TQ	Q	FL	P	FS	QF	C	LY	D	GS	LS	S
Niers	terraces																	
CX22-1	Floodplain	193,35	416,55	300	37.7	1	9.2	47.9	0	1.3	0.7	0.7	0.3	0.3	7.2	41.3	0	48.5
CB5a-3	Gennep	193,95	414,50	300	37.6	1	13.8	52.4	0	0.6	0.6	1.6	0	0.3	6.1	38.3	0	44.4
CB9-3	Gennep	194,30	415,14	172	36.4	0.3	9.3	46.1	1.2	0.9	1.2	1.2	0.3	0.9	11.1	37.3	0	48.4
CB1-18	Gennep	193,55	414,23	305	39.4	2.5	14.9	56.8	0	1	0.5	1	0	0.3	9.8	30.6	0	40.4
CE6-1	Gennep	196,93	412,78	300	39.3	6	19	64.3	0	0	0.7	1	0	0.3	2.7	18.3	0	31
CE7-1	Gennep	197,00	412,70	276	36.2	2.2	25.7	64.1	0.7	0.7	1.5	1.8	0	1.1	2.6	20.7	0	27.7
CB11-1	Milsbeek	194,80	415,75	300	39.7	1.8	17	58.5	1	0.5	0.5	1	0	0.8	10.3	27.3	0	37.7
CX24-1	Milsbeek	193,35	416,41	296	33.8	5.1	15.9	54.7	0.3	0.7	0.3	0	0	0.3	8.8	34.8	0	43.6
CX3-1	Milsbeek	193,00	416,80	300	33.7	2	19.3	55	1	1.7	0.3	1.7	0	0.3	9.7	30.3	0	40
CX4-1	Milsbeek	193,13	416,75	300	43.7	0.7	18.3	62.7	0.3	1	0	0	0	0.3	8	27.7	0	35.7
CB10-1	Milsbeek	194,60	415,43	324	39.1	3.2	16.6	58.9	0.6	1.9	0.5	0.3	0.5	0.3	7.6	29.4	0	37
CX21-3	Milsbeek	192,56	415,20	300	52.7	1	11.7	65.3	0	0	0.3	0	0	0.3	5.3	28.7	0	34
Maas	terraces																	
C22-3	floodplain	186,79	416,33	300	47.7	2.3	14.3	64.3	0	1	1	1.3	0	0	6.3	26	0	32.3
CB19-1	floodplain	192,83	413,60	300	48.3	2.7	14.5	65.5	0.7	0.7	0.2	0.3	0	0.3	7.5	24.8	0	32.3
C14-2	floodplain	185,45	418,43	310	47.3	0.8	13.3	61.4	0.3	1.3	0.2	1.5	0.2	0.2	4.2	30.8	0	35
CB14-2	Wanssum	192,83	413,33	260	43.1	1.2	15.4	59.6	0.4	1.5	1.2	0	0	0	12.3	24.6	0.4	36.9
C8-1	Wanssum	187,66	416,79	300	45	4	15.3	64.3	0.3	0.7	0.3	0	0	0	7.3	27	0	34.3
RW-1	Vierlingsbeek	185,58	413,45	203	50.3	2.5	17.2	70	0	0	0	0.5	0	0.5	3.5	25.6	0	29.1
C25-1	Vierlingsbeek	185,10	413,58	187	55.6	2.7	20.9	79.2	1.6	0	0	0.5	0	0	0.5	18.2	0	18.7
C19-3	Vierlingsbeek	186,25	414,29	252	44.4	4	11.5	59.9	1.2	2	0.4	1.6	2.8	0	14.3	17.5	0.4	31.7
CB24-2	Vierlingsbeek	191,96	412,34	309	35.4	2.4	23.9	61.7	1	1	0.7	1.6	0	0.2	8.5	25.5	0	33.9
C2-5	Rijkevoort	183,53	409,70	218	39.9	0.5	23.2	63.5	0.5	0.5	2.1	0.5	0	0	4.3	28.7	0	33
C24-2	Overloon	183,10	409,00	300	55.8	1.2	10.8	67.8	0.8	0.3	1.2	2.8	0	0.3	5.2	21.5	0	26.7
CB5-30	riverdune	193,95	414,50	300	42	1.3	24.7	68	0	0.7	0.3	0.7	0	0	5.7	24.7	0	30.3

Table 2. Heavy mineral composition of Weichselian and Holocene deposits in the Maas and Niers valleys (this study), in percentages. Size class 2= 75-105 mu, 3= 105-150 mu, 4= 150-210 mu, 5= 210-300 mu, 6= 300-420 mu; n= number of grains counted; G= garnet group; E= epidote group; A= alterite group; H= hornblende group; C= chloritoid group; V= volcanic group; S= stable group; U= unstable group; TS= topaz/staurolite group; M= metamorphic group; T= tourmaline; O= opaque minerals; holo dune= Holocene dune; YD riverdune= Younger Dryas riverdune.

sample		x-coord	y-coord	size class	n	G	E	A	H	C	V	S	U	TS	M	T	O
NIERS	terraces																
cb1-8	floodplain	193.55	414.22	6	197	27.7	14.9	12.4	9.4	2.0	7.6	11.0	3.0	4.5	2.5	5.4	30
cb9-1	floodplain	194.20	416.12	4	199	41.0	14.0	7.5	10.6	1.0	6.5	7.5	0.5	5.5	3.5	3.0	30
zm-26	floodplain	191.38	415.08	silt	213	22.1	16.9	13.1	5.6	0.0	8.0	17.4	1.9	6.1	1.9	7.0	38
zm-14	Gennep	198.48	413.25	4	210	11.9	24.3	26.2	14.8	0.5	5.7	2.4	0.0	1.4	4.8	8.1	77
b55-1	Gennep	198.40	409.15	5	101	14.9	16.8	17.8	5.9	1.0	8.9	10.9	2.0	7.9	5.9	7.9	61
ce11-1	Milsbeek	197.09	415.26	5	210	33.8	12.4	18.6	10.5	0.0	8.1	4.8	0.5	5.2	3.3	2.9	30
NIERS MAAS confluen	ce area																
c12-2	floodplain	184.60	411.38	3	200	32.5	15.0	16.5	7.0	0.5	3.5	6.0	3.0	5.0	2.0	9.0	51
c14-1	floodplain	185.45	418.43	3	200	9.5	18.0	5.5	7.5	0.0	2.0	29.0	1.5	8.0	3.5	15.5	72
c22-1	floodplain	186.79	416.33	4	200	21.0	21.5	6.5	5.5	1.0	3.5	14.5	1.0	6.5	5.5	13.5	61
c22-2	floodplain	186.79	416.33	2	200	2.0	19.0	7.5	2.5	1.0	1.0	46.0	4.0	2.5	4.5	10.0	58
cb14-1	Wanssum	192.83	413.33	4	172	19.5	21.0	11.0	5.5	1.5	2.0	18.0	1.5	3.0	8.0	9.0	63
c8-1	Wanssum	187.66	416.79	6	200	18.5	8.0	8.0	10.5	1.5	27.0	8.5	0.0	2.0	7.5	8.5	44
c15-1	Broekhuizen	187.24	418.03	3	200	12.0	19.0	9.0	15.5	0.0	4.0	21.0	1.0	5.0	6.0	7.5	61
c15-3	Broekhuizen	187.24	418.03	4	200	16.5	12.0	17.0	8.0	0.0	5.0	20.0	2.5	4.0	6.0	9.0	67
zm-16	Vierlingsbeek	197.35	413.88	4	211	13.3	23.7	20.9	17.1	0.5	4.3	6.6	0.9	3.3	3.8	5.7	26
zm-24	Vierlingsbeek	196.15	413.25	4	195	15.9	16.4	28.7	14.4	0.0	10.3	4.1	1.5	1.0	4.1	3.6	38
cb1-16	Vierlingsbeek	192.46	413.14	6	191	36.0	6.0	9.5	12.5	0.5	12.1	6.1	6.0	8.0	1.0	2.5	46
cb3-2	Vierlingsbeek	193.77	414.42	4	193	33.5	16.5	9.0	9.0	1.0	9.0	6.0	4.0	4.0	4.0	4.0	51
cb5a-2	Vierlingsbeek	193.95	414.50	4	196	41.5	13.0	11.0	4.0	3.5	10.6	8.0	1.0	1.5	3.5	2.5	54
cb24-1	Vierlingsbeek	191.96	412.34	3	157	21.5	13.0	15.5	4.5	0.0	4.5	10.5	2.5	9.5	3.0	15.5	60
cb28-3	Vierlingsbeek	194.28	414.60	4	236	29.5	18.5	14.5	8.5	0.0	9.0	5.0	3.5	4.0	5.0	2.5	43
cb7-2	Vierlingsbeek	193.67	414.82	4	201	26.5	20.5	10.0	12.5	0.5	10.4	6.5	4.0	3.0	3.5	2.5	40
cx20-1	Vierlingsbeek	193.03	415.08	4	192	26.6	16.7	7.8	6.3	0.0	4.2	21.4	0.5	5.2	3.6	7.8	36
c19-2	Vierlingsbeek	186.25	414.29	4	201	23.0	12.5	7.5	5.5	0.5	6.5	25.0	5.0	3.0	3.0	9.0	42
c20-1	Vierlingsbeek	186.59	414.84	3	200	7.0	18.0	13.5	7.5	1.5	4.5	26.5	2.0	4.0	5.0	10.5	40
c21-3	Vierlingsbeek	186.80	415.74	3	200	12.5	12.5	19.5	3.0	1.5	8.5	15.0	2.5	5.0	4.5	15.5	46
x8-1	Vierlingsbeek	181.32	414.10	5	123	16.3	12.2	30.1	4.1	1.6	17.9	0.8	0.8	1.6	3.3	11.4	23
rw-2	Vierlingsbeek	185.63	413.53	5	200	31.5	15.0	20.5	2.0	1.5	3.0	8.0	2.5	5.0	0.5	10.5	27
zm-54	Vierlingsbeek	193.30	408.40	5	211	7.1	14.7	27.5	10.0	0.5	6.6	11.4	1.4	5.7	6.2	9.0	74
cx24-2	Rijkevoort	193.35	416.45	3	213	23.5	23.0	10.3	11.7	0.0	8.9	6.6	3.8	2.8	3.3	6.1	29
zm-3	Rijkevoort	193.70	416.18	4-5	212	4.7	6.6	46.7	11.3	0.0	20.3	0.5	0.0	3.8	3.3	2.8	25
zm-5	Rijkevoort	194.25	416.53	3	209	14.8	15.3	37.3	11.5	0.0	6.7	1.9	3.8	2.9	3.3	2.4	40
zm-6	Rijkevoort	195.90	416.25	4	222	16.7	18.5	25.7	12.2	0.5	6.3	2.7	0.9	4.1	3.6	9.0	59
zm-7	Rijkevoort	196.93	416.65	4	212	17.5	15.1	37.7	7.5	0.0	6.1	1.9	3.3	3.8	4.7	2.4	43
zm-10	Rijkevoort	198.45	415.58	3-4	257	12.1	22.2	17.1	12.1	0.0	3.5	5.4	1.2	8.9	4.7	12.8	38
c13-1	Rijkevoort	185.05	412.10	3	200	33.0	13.5	11.5	3.0	0.0	2.0	15.5	2.0	6.5	4.5	8.5	32
c18-3	Rijkevoort	185.38	412.70	3	201	57.5	8.5	8.0	1.5	0.0	2.5	8.5	3.5	3.5	3.0	4.0	18
c23-2	Rijkevoort	184.78	411.79	4	200	12.5	19.5	33.0	7.0	3.0	9.0	2.0	0.0	4.5	2.0	7.5	17
aeolian	*deposits*																

cb5-9	holo dune	193.95	414.50	5	195	33.5	11.6	15.5	7.5	2.1	9.6	2.6	1.5	3.0	6.5	7.0	48
cb5-15	YDriverdune	193.95	414.50	5	191	28.0	12.5	14.0	11.5	2.0	8.5	10.0	3.5	4.0	4.0	2.0	45
cb5-20	YDriverdune	193.95	414.50	5	186	22.0	12.5	22.0	8.5	2.0	13.0	3.0	2.5	2.0	5.0	7.5	46
c26-1	coversand	183.73	410.00	2	200	37.5	18.5	7.5	0.0	0.5	1.0	19.0	3.0	4.0	3.5	5.5	25
MAAS	terraces																
cb13-2	floodplain	193.35	413.82	5	133	14.9	8.5	12.8	11.2	2.1	10.2	2.1	1.6	12.3	6.4	17.6	38
cf4-1	floodplain	194.18	407.75	3	195	16.4	16.9	14.4	6.7	1.0	5.6	7.7	2.6	9.7	3.1	15.9	60
32.20	floodplain	209.80	386.80	3	100	7.0	24.0	1.0	1.0	0.0	3.0	47.0	0.0	4.0	6.0	7.0	67
m13-1	floodplain	199.80	399.40	4	100	14.0	10.0	4.0	3.0	2.0	1.0	46.0	0.0	3.0	3.0	14.0	61
w9-2	floodplain	204.90	395.50	3	93	21.5	7.5	5.4	4.3	4.3	3.2	37.6	0.0	3.2	5.4	7.5	70
w15-2	floodplain	203.30	392.70	4	91	17.6	17.6	1.1	9.9	1.1	3.3	14.3	0.0	9.9	8.8	16.5	73
x3-1	local creek	181.36	414.16	4	201	37.5	15.0	12.0	6.5	1.0	4.0	8.5	2.5	4.0	4.5	5.0	23
x15-2	local creek	187.90	409.70	5	200	31.5	18.0	9.5	5.5	0.5	5.5	16.0	1.0	4.5	5.0	3.0	32
x15-4	local creek	187.90	409.70	4	201	22.0	18.5	17.0	6.0	1.0	3.5	5.5	1.0	9.5	7.0	9.5	14
x16-2	local creek	204.25	393.95	4	200	35.0	13.0	22.0	6.0	0.5	5.0	2.5	0.5	4.5	4.0	7.0	21
lv3-2	Wanssum	198.18	403.04	4	199	8.0	13.1	6.0	0.5	1.0	4.0	42.7	0.5	6.0	9.5	8.5	49
sb-20	Wanssum	200.93	399.40	6	202	29.7	19.3	5.0	1.5	0.5	1.5	12.4	1.0	10.9	4.0	14.4	42
w15-3	Wanssum	203.30	392.70	2	100	20.0	20.0	8.0	7.0	1.0	1.0	14.0	0.0	8.0	6.0	15.0	43
cb21-2	Broekhuizen	192.41	412.30	3	178	29.0	15.5	9.0	2.0	1.5	3.0	20.5	1.0	4.5	4.0	10.0	59
44.60	Broekhuizen	207.80	387.20	3	100	7.0	20.0	1.0	3.0	0.0	4.0	44.0	0.0	4.0	3.0	14.0	69
44.10	Broekhuizen	207.80	387.20	4	100	7.0	19.0	1.0	1.0	0.0	5.0	34.0	0.0	13.0	6.0	14.0	69
41.20	Broekhuizen	210.20	385.40	2	100	4.0	18.0	2.0	3.0	1.0	5.0	42.0	0.0	4.0	5.0	16.0	66
b53-1	Broekhuizen	196.60	408.40	4	101	12.9	9.9	6.9	2.0	0.0	4.0	39.6	1.0	10.9	4.0	8.9	68
b29-1	Broekhuizen	192.26	410.60	5	100	13.0	6.0	14.0	2.0	2.0	9.0	18.0	0.0	13.0	10.0	13.0	76
b31-1	Broekhuizen	192.63	410.74	6	99	14.1	12.1	11.1	4.0	1.0	6.1	23.2	2.0	6.1	4.0	16.2	58
zm-60	Vierlingsbeek	194.73	405.33	4	202	9.4	16.8	38.6	9.9	0.0	4.0	4.0	5.9	7.9	1.0	2.5	46
zm-61	Vierlingsbeek	195.53	404.53	4	183	13.7	17.5	18.6	12.0	0.0	8.2	12.6	4.4	4.4	4.4	4.4	51
zm-62	Vierlingsbeek	197.43	401.88	4	146	5.5	13.0	19.9	10.3	0.0	3.4	18.5	2.1	4.1	11.0	12.3	63
bw-2m1	Vierlingsbeek	198.60	397.94	4	202	5.0	25.7	8.4	2.5	1.5	2.0	17.3	5.0	11.9	5.9	14.9	52
hh-10	Vierlingsbeek	197.85	398.32	4	201	30.3	12.4	4.5	2.5	0.0	1.5	16.9	2.5	7.0	7.0	15.4	34
hh-12	Vierlingsbeek	197.85	398.32	3	200	27.0	14.0	3.0	3.0	1.0	3.5	22.5	3.0	4.5	7.0	11.5	34
m12-4	Vierlingsbeek	199.90	399.00	5	98	15.3	9.2	9.2	3.1	0.0	2.0	32.7	2.0	8.2	3.1	15.3	61
m3-2	Vierlingsbeek	199.10	398.00	5	100	6.0	17.0	9.0	2.0	0.0	10.0	26.0	2.0	11.0	8.0	9.0	37
m12-1	Vierlingsbeek	199.90	399.00	3	101	5.0	15.8	4.0	4.0	2.0	1.0	37.6	1.0	6.9	5.9	16.8	66
w1a-1	Vierlingsbeek	202.80	392.50	4	100	19.0	11.0	4.0	2.0	0.0	0.0	50.0	1.0	2.0	5.0	6.0	43
w1a-2	Vierlingsbeek	202.80	392.50	4	101	29.7	6.9	5.0	5.9	0.0	1.0	29.7	1.0	5.0	7.9	7.9	44
g3-3	Vierlingsbeek	199.50	394.20	5	100	13.0	10.0	8.0	1.0	1.0	2.0	19.0	1.0	13.0	10.0	22.0	37
zm-59	Rijkevoort	193.68	404.65	4	225	27.6	14.2	20.4	8.0	0.0	3.1	4.4	0.4	7.1	5.8	8.9	58
zm-63	Rijkevoort	196.25	400.70	4-5	185	33.0	16.8	13.5	4.9	1.1	3.2	17.8	1.6	1.6	3.8	2.7	54
zm-64	Rijkevoort	195.90	401.63	4	172	22.1	19.2	9.9	6.4	1.2	2.3	33.1	0.0	1.7	2.9	1.2	22
zm-67	Rijkevoort	191.78	406.08	4	205	31.7	10.7	16.1	6.3	0.0	3.4	15.1	2.4	2.0	4.9	7.3	46
lb-17	Rijkevoort	197.05	397.70	5	204	13.2	8.8	6.9	5.4	0.0	2.9	23.0	2.5	9.3	8.8	19.1	34
bhbp7	Rijkevoort	204.25	398.35	3	100	19.0	18.0	8.0	5.0	0.0	1.0	31.0	0.0	4.0	5.0	9.0	42
bh6	Rijkevoort	204.60	398.60	4	101	28.7	13.9	12.9	5.9	3.0	2.0	13.9	1.0	4.0	5.9	8.9	35
bh3	Rijkevoort	204.60	398.60	4	100	22.0	14.0	9.0	6.0	1.0	3.0	21.0	1.0	8.0	8.0	7.0	38
bh7	Rijkevoort	204.60	398.60	2	100	22.0	15.0	10.0	12.0	1.0	3.0	21.0	0.0	4.0	4.0	8.0	35
g5-2	Rijkevoort	199.80	393.80	6	100	29.0	3.0	7.0	2.0	0.0	4.0	32.0	1.0	6.0	7.0	9.0	33
o1	Rijkevoort	199.60	396.20	6	105	7.6	8.6	17.1	4.8	1.9	6.7	21.0	2.9	8.6	5.7	15.2	41

sample		x-coord	y-coord	size class	n	G	E	A	H	C	V	S	U	TS	M	T	O
b1-1	Rijkevoort	187.90	409.70	5	101	34.7	9.9	11.9	1.0	2.0	0.0	20.8	3.0	7.9	3.0	5.9	37
b2-2	Rijkevoort	188.90	409.60	4	101	31.7	16.8	4.0	4.0	0.0	2.0	24.8	2.0	9.9	1.0	4.0	42
cx7-2	Rijkevoort	189.50	407.28	4	200	41.5	9.5	14.0	4.5	0.5	2.0	10.5	0.0	5.5	4.0	8.0	23
cx7-1	Rijkevoort	189.50	407.28	4	200	30.0	11.5	18.5	6.5	1.5	1.5	13.0	1.5	2.0	5.0	9.0	23
cx7-4	Rijkevoort	189.50	407.28	4	200	37.0	14.5	7.0	2.5	0.5	1.0	27.5	0.5	2.5	2.5	4.5	33
c2-5	Rijkevoort	183.53	409.70	3	200	15.0	15.0	10.5	7.5	0.0	4.5	13.0	1.0	10.5	6.0	17.0	36
bh94-6	Rijkevoort	204.70	398.40	5	101	26.7	8.9	7.9	5.9	0.0	6.9	24.8	2.0	4.0	5.9	6.9	25
bh94-12	Rijkevoort	204.70	398.40	5	100	29.0	21.0	13.0	5.0	2.0	4.0	8.0	0.0	5.0	2.0	11.0	26
bh94-7	Rijkevoort	204.70	398.40	4	100	23.0	15.0	19.0	7.0	3.0	1.0	10.0	0.0	4.0	7.0	11.0	25
bh94-8	Rijkevoort	204.70	398.40	6	100	15.0	6.0	32.0	13.0	0.0	23.0	5.0	0.0	2.0	3.0	1.0	11
bh94-9	Rijkevoort	204.70	398.40	3	100	22.0	20.0	12.0	7.0	1.0	2.0	15.0	0.0	6.0	8.0	7.0	30
bh94-1	Rijkevoort	204.65	398.35	4	100	35.0	12.0	12.0	4.0	1.0	0.0	13.0	1.0	5.0	5.0	12.0	22
bh94-13	Rijkevoort	204.65	398.35	4	100	27.0	15.0	8.0	4.0	0.0	2.0	29.0	2.0	2.0	4.0	7.0	40
bh94-14	Rijkevoort	204.20	398.33	5	100	14.0	23.0	15.0	5.0	0.0	1.0	22.0	1.0	3.0	7.0	9.0	29
bh94-15	Rijkevoort	204.20	398.33	4	100	27.0	15.0	10.0	1.0	0.0	1.0	27.0	2.0	4.0	2.0	11.0	34
bh94-16	Rijkevoort	204.20	398.33	4	100	20.0	14.0	18.0	6.0	0.0	6.0	18.0	0.0	6.0	5.0	7.0	26
bh94-17	Rijkevoort	204.20	398.33	5	100	24.0	17.0	19.0	4.0	0.0	2.0	15.0	0.0	5.0	1.0	13.0	26
bh94-18	Rijkevoort	204.20	398.33	6	100	29.0	9.0	21.0	7.0	0.0	7.0	14.0	0.0	0.0	5.0	8.0	22
deursen	Overloon	171.20	423.93	4		30.0	18.6	3.8	1.9	0.5	5.2	22.9	1.4	2.9	6.2	6.7	48
c1-4	Overloon	183.15	409.38	5	163	8.0	8.5	9.0	3.5	0.0	1.0	3.5	0.5	12.0	4.0	31.5	31
c24-1	Overloon	183.10	409.00	4	152	15.1	8.6	7.2	4.6	0.7	2.6	14.5	1.3	13.8	7.2	24.3	42
aeolian	*deposits*																
bh24	holo dune	203.90	398.42	5	100	25.0	14.0	9.0	1.0	0.0	1.0	15.0	1.0	13.0	6.0	15.0	75
bhbp2	YDriverdune	204.25	398.35	4	100	26.0	12.0	14.0	4.0	0.0	1.0	21.0	2.0	6.0	4.0	10.0	31
bhbp1	YDriverdune	204.25	398.35	5	100	24.0	13.0	7.0	1.0	0.0	0.0	38.0	0.0	4.0	0.0	13.0	44
bh6-18	YDriverdune	204.60	398.60	5	100	29.0	6.0	8.0	1.0	1.0	1.0	25.0	0.0	10.0	6.0	13.0	46
cx7-3	coversand	189.50	407.28	3	200	41.5	10.5	9.5	1.0	0.0	1.0	21.5	2.0	4.5	4.5	4.0	27
c2-1	coversand	183.53	409.70	3	200	21.5	23.5	11.5	4.0	0.0	3.0	9.5	2.5	8.5	4.0	12.0	21
c2-3	coversand	183.53	409.70	3	200	4.0	17.5	6.5	4.5	1.0	1.0	29.5	2.5	10.0	8.0	15.5	51
c5-2	coversand	184.36	411.01	4	200	49.5	10.0	8.0	3.0	0.0	1.5	13.5	1.0	4.0	2.0	7.5	22
DIVERSE																	
lb-32	preWeichsel	197.05	397.70	4	200	8.5	9.0	7.0	10.5	1.0	9.0	14.0	1.0	10.5	11.5	18.0	32
sb-34	preWeichsel	200.93	399.40	4	202	16.3	15.3	15.8	0.5	2.5	0.5	19.8	0.0	10.9	2.0	16.3	34
lv1-2	preWeichsel	199.00	403.00	4	103	11.7	7.8	24.3	3.9	1.0	1.9	12.6	5.8	9.7	4.9	16.5	33

Table 3. Heavy mineral composition of Weichselian and Holocene deposits in the Maas and Niers valleys (analysis by the RGD = State Geological Survey of The Netherlands), in percentages. Legend see table 2.

sample		x-coord	y-coord	size class	n	G	E	A	H	C	V	S	U	S	M	T	O
NIERS	MAASconfluen	ce area															
JISZ	Wanssum	187.85	417.05	4-5	100	18	23	5	10	2	7	19	0	5	4	7	24
29542	Vierlingsbeek	187.50	415.94	>3	200	18	17.5	14.5	5.5	2	7	10.5	0	9	3.5	12.5	
29543	Vierlingsbeek	187.50	415.94	>3	200	25	11.5	14.5	4	0	13.5	9.5	0	7	5	10	
31356	Vierlingsbeek	190.92	410.4	4	200	6	27	25.5	6.5	2	10	4	0	4.5	5	9.5	
31357	Vierlingsbeek	190.92	410.4	4	200	17.5	20	25.5	12	0	3.5	10	0	4	2.5	4.5	
11072	Vierlingsbeek	196.35	410.15	>3	100	8	9	33	7	0	14	11	0	6	2	10	
11073	Vierlingsbeek	196.35	410.15	>3	100	5	4	26	5	0	41	2	0	4	3	10	
11074	Vierlingsbeek	196.35	410.15	>3	100	4	21	38	5	0	7	3	0	6	1	15	
2510	Vierlingsbeek	196.41	409.73	>3	100	15	33	12	5	0	11	10	0	6	1	7	37
2511	Vierlingsbeek	196.41	409.73	>3	100	6	8	26	14	0	18	0	0	4	9	16	50
2512	Vierlingsbeek	196.41	409.73	>3	100	28	10	20	4	0	19	7	0	3	1	7	21
2513	Vierlingsbeek	196.41	409.73	>3	100	23	27	9	6	0	4	8	0	10	3	10	46
2514	Vierlingsbeek	196.41	409.73	>3	100	36	19	9	3	0	8	7	0	4	3	11	41
2515	Vierlingsbeek	196.41	409.73	>3	100	16	9	36	6	0	17	1	0	6	4	5	21
2516	Vierlingsbeek	196.41	409.73	>3	100	28	10	33	8	0	9	0	0	5	4	3	16
JISZ	Rijkevoort	187.53	410.21	3?	100	24	14	19	8	0	12	1	0	7	8	7	11
JISZ	Rijkevoort	187.53	410.21	3?	100	32	11	15	5	0	11	4	0	4	8	10	
JISZ	Rijkevoort	187.53	410.21	3?	100	24	17	20	4	0	19	1	0	7	2	6	13
17323	Rijkevoort	202.85	403.74	>3	100	27	12	26	11	0	15	3	0	4	1	1	
17324	Rijkevoort	202.85	403.74	>3	100	26	20	24	16	0	4	3	0	3	1	3	
17325	Rijkevoort	202.85	403.74	>3	100	11	12	34	15	0	14	0	0	2	3	9	
30969	Rijkevoort	192.92	405.7	>3	200	42.5	16	13	5.5	0	1	13.5	0	3	1	4.5	
30970	Rijkevoort	192.92	405.7	>3	200	39.5	12	13	6	0.5	1	11.5	0	5	1	10.5	
1571	Rijkevoort	202.15	404.2	>3	100	7	8	36	12	0	30	0	0	1	2	4	
1572	Rijkevoort	202.15	404.2	>3	100	13	8	49	6	0	14	1	0	2	3	4	
5372	Rijkevoort	200.06	405.81	>3	100	4	9	26	10	0	30	1	0	6	7	7	
5373	Rijkevoort	200.06	405.81	>3	100	29	9	21	6	0	5	15	0	3	4	8	
5374	Rijkevoort	200.06	405.81	>3	100	15	22	29	6	0	15	4	0	2	2	5	
5375	Rijkevoort	200.06	405.81	>3	100	20	18	32	4	2	5	5	0	5	2	7	
5376	Rijkevoort	200.06	405.81	>3	100	8	18	40	9	0	9	2	0	2	5	7	
17322	Rijkevoort	202.85	403.74	>3	100	11	23	34	16	0	10	2	0	2	1	1	
MAAS	terraces																
16834	floodplain	200.10	400.15	>3	100	12	7	20	16	0	4	11	0	6	7	17	
16835	floodplain	200.10	400.15	>3	100	18	8	8	14	0	5	18	0	15	4	10	
16836	floodplain	200.10	400.15	>3	100	20	8	17	8	0	4	19	0	16	1	7	
12211	floodplain	205.24	395.31	3	100	13	12	1	9	0	0	46	0	3	4	12	
12212	floodplain	205.24	395.31	3	100	23	8	2	1	0	0	50	0	5	2	9	
9176	Wanssum	201.65	395.35	>3	100	13	5	6	11	0	5	9	0	15	11	25	
11025	Wanssum	201.93	397.40	>3	100	17	8	10	1	0	6	4	0	19	19	16	
11026	Wanssum	201.93	397.40	>3	50	14	10	16	0	0	4	0	0	12	10	34	
11027	Wanssum	201.93	397.40	>3	64	19	6	14	3	0	5	0	0	8	9	36	
9768	Wanssum	202.23	394.52	>3	100	14	26	16	12	0	0	10	0	8	4	10	

sample		x-coord	y-coord	size class	n	G	E	A	H	C	V	S	U	S	M	T	O
5380	Wanssum	202.46	397.46	>3	100	11	13	28	10	1	18	1	0	9	3	6	
5381	Wanssum	202.46	397.46	>3	100	13	6	14	4	3	0	6	0	12	12	30	
9130	Wanssum	204.15	394.03	3	100	5	9	10	4	0	0	23	0	16	9	24	
9131	Wanssum	204.15	394.03	3	100	9	8	11	9	0	1	15	0	20	7	20	
9132	Wanssum	204.15	394.03	3	100	4	10	10	11	0	0	27	0	11	9	18	
9133	Wanssum	204.15	394.03	3	100	4	14	8	5	0	0	11	0	18	8	32	
12244	Wanssum	203.54	394.56	>3	100	19	13	6	1	0	0	22	0	14	10	15	
12245	Wanssum	203.54	394.56	>3	100	17	13	8	5	0	1	35	0	5	9	7	
12246	Wanssum	203.54	394.56	>3	100	6	17	11	1	0	3	9	0	13	10	30	
34072	Broekhuizen	208.12	391.58	>3	200	15	5.5	5.5	2	4.5	1	32.5	0	9	8	17	
34073	Broekhuizen	208.12	391.58	>3	200	13.5	6.5	3.5	0.5	4	1	9.5	0	15	8	38.5	
28985	Broekhuizen	204.65	393.23	3?	200	26	9	4.5	3	1	1.5	36	0	4	5	10	
12191	Broekhuizen	200.52	392.95	>3	100	2	19	21	4	0	0	19	0	9	9	17	
12192	Broekhuizen	200.52	392.95	>3	100	5	11	25	10	0	1	14	0	10	7	18	
5333	Broekhuizen	208.47	390.35	3	100	2	18	3	3	6	0	28	0	15	8	17	
5334	Broekhuizen	208.47	390.35	3	100	14	11	4	2	9	0	28	0	13	6	13	
29002	Broekhuizen	386.64	207.57	>3	200	24.5	9	2.5	2.5	0	0.5	18.5	0	13.5	8	21	
29003	Broekhuizen	386.64	207.57	>3	200	27.5	6	2	7.5	0	0.5	11.5	0	14	9	22	
29004	Broekhuizen	386.64	207.57	>3	200	29.5	5	4	3	1	0	9	0	17	6	25.5	
9695	Vierlingsbeek	201.43	394.51	>3	100	8	19	15	7	0	1	17	0	11	6	16	
9691	Vierlingsbeek	201.43	394.51	>3	100	18	22	3	10	0	1	18	0	8	4	16	
9642	Vierlingsbeek	201.43	394.51	>3	100	27	15	4	3	0	0	23	0	12	6	11	
9878	Vierlingsbeek	202.60	393.9	3	100	10	18	3	7	0	2	29	0	9	6	16	
9881	Vierlingsbeek	202.60	393.9	3	100	12	21	6	12	0	1	27	0	6	4	11	
9982	Vierlingsbeek	202.60	393.9	3	100	17	19	9	7	0	1	28	0	9	3	7	
9806	Vierlingsbeek	202.08	394.5	3	100	10	21	11	2	0	0	26	0	12	4	14	
9807	Vierlingsbeek	202.08	394.5	3	100	11	20	14	7	0	2	14	0	11	6	15	
9808	Vierlingsbeek	202.08	394.5	3	100	4	17	10	14	0	2	17	0	12	8	15	
9809	Vierlingsbeek	202.08	394.5	3	100	5	15	21	7	0	3	3	0	8	11	27	
9811	Vierlingsbeek	202.08	394.5	3	100	7	22	6	9	0	0	27	0	7	5	17	
9316	Vierlingsbeek	202.36	393.95	3	100	6	13	9	6	0	0	43	0	3	7	13	
9801	Vierlingsbeek	202.78	393.76	3	100	11	12	6	6	0	0	26	0	13	12	14	
9802	Vierlingsbeek	202.78	393.76	3	100	9	12	10	8	0	1	21	0	13	13	13	
9803	Vierlingsbeek	202.78	393.76	3	100	9	20	16	5	0	1	22	0	13	5	9	
9805	Vierlingsbeek	202.78	393.76	3	100	13	9	14	9	0	0	11	0	9	14	20	
14991	Vierlingsbeek	194.15	404.52	4-5	100	13	16	36	14	0	6	6	0	3	2	4	
14992	Vierlingsbeek	194.15	404.52	4-5	100	16	19	39	11	0	1	4	0	4	2	4	
14993	Vierlingsbeek	194.15	404.52	4-5	100	18	20	30	6	0	1	5	0	6	4	10	
9797	Vierlingsbeek	201.40	395.1	>3	100	17	12	10	10	0	2	11	0	19	6	13	
9748	Vierlingsbeek	201.43	394.51	>3	100	2	21	10	14	0	0	15	0	14	9	15	
5341	Rijkevoort	207.83	394.85	>3	100	13	21	8	3	3	0	19	0	8	8	17	
5342	Rijkevoort	207.83	394.85	>3	100	21	11	7	3	0	0	3	0	20	6	29	
5367	Rijkevoort	208.80	393.83	>3	100	34	13	8	3	1	1	12	0	13	10	5	
5368	Rijkevoort	208.80	393.83	>3	100	22	13	5	3	3	0	5	0	10	11	28	
5393	Rijkevoort	208.01	393.45	>3	100	11	11	8	2	0	2	5	0	20	7	34	
5394	Rijkevoort	208.01	393.45	>3	100	8	20	45	8	1	9	1	0	1	4	3	

sample		x-coord	y-coord	size class	n	G	E	A	H	C	V	S	U	S	M	T	O
5395	Rijkevoort	208.01	393.45	>3	100	7	19	8	3	2	1	10	0	11	7	32	
5156	Rijkevoort	210.85	388.65	>3	100	10	11	23	5	0	1	12	0	7	7	24	
5157	Rijkevoort	210.85	388.65	>3	100	17	5	6	7	0	1	5	0	13	11	35	
3192	Rijkevoort	211.70	387.98	>3	100	26	17	5	3	0	0	28	0	13	2	6	
5168	Rijkevoort	211.70	387.98	>3	100	28	10	2	0	0	0	8	0	24	8	19	
10977	Rijkevoort	210.50	391.36	>3	100	36	8	12	3	0	2	4	0	10	8	16	
10978	Rijkevoort	210.50	391.36	>3	100	29	19	7	4	0	0	7	0	13	5	16	
5202	Rijkevoort	204.50	397.51	>5	100	29	10	8	4	0	5	13	0	9	6	16	
5203	Rijkevoort	204.50	397.51	>5	100	17	8	12	10	0	15	3	0	15	5	15	
5204	Rijkevoort	204.50	397.51	>5	100	23	7	8	3	1	0	2	0	14	8	33	
8817	Rijkevoort	203.30	398.13	>3	100	34	12	3	4	0	8	20	0	7	3	9	
8818	Rijkevoort	203.30	398.13	>3	100	40	19	17	6	0	5	1	0	6	0	6	
5226	Rijkevoort	203.65	398.81	3	100	18	23	12	4	0	4	11	0	4	5	19	
5227	Rijkevoort	203.65	398.81	>3	100	12	15	18	6	0	18	8	0	9	5	9	
5228	Rijkevoort	203.65	398.81	>3	100	19	19	27	3	0	22	0	0	2	2	6	
5347	Rijkevoort	203.65	398.81	>3	100	15	8	24	5	0	26	1	0	6	6	9	
5230	Rijkevoort	203.65	398.81	>3	100	25	11	5	5	1	0	7	0	9	13	24	
5211	Rijkevoort	203.06	399.95	>3	100	30	11	9	1	0	2	25	0	10	3	8	
5212	Rijkevoort	203.06	399.95	>3	100	25	11	23	5	0	9	7	0	4	1	15	
25260	Rijkevoort	204.81	400.21	>3	200	9	9	37.5	11.5	0	29	0.5	0	1.5	0	1.5	
9084	Rijkevoort	201.81	393.6	3	100	23	20	10	2	0	0	28	0	4	2	11	
9085	Rijkevoort	201.81	393.6	3	100	15	11	8	3	0	0	39	0	8	3	13	
9086	Rijkevoort	201.81	393.6	3	100	7	26	25	6	0	1	5	0	9	5	16	
5142	Rijkevoort	202.71	401.49	3	100	3	11	43	16	0	21	1	0	2	1	2	
5143	Rijikevoort	202.71	401.49	3	100	22	11	40	13	0	6	0	0	4	0	4	
5184	Rijikevoort	205.61	400.1	>3	100	12	13	41	8	0	9	4	0	4	5	4	
5185	Rijikevoort	205.61	400.1	>3	100	0	18	39	4	0	7	25	0	4	1	2	
25268	Rijikevoort	204.10	400.00	>3	200	24	14.5	18.5	7.5	0	3.5	10	0	7	2.5	12.5	
25269	Rijikevoort	204.10	400.00	>3	200	7	5.5	34	13	0	37	0	0	1	2	0.5	
25270	Rijikevoort	204.10	400.00	>3	200	32	16.5	15.5	4	0	8.5	2.5	0	8	5.5	7.5	
25271	Rijikevoort	204.10	400.00	>3	200	11	7.5	20	6.5	0	47	1	0	2.5	2	2.5	
25272	Rijikevoort	204.10	400.00	>3	200	14.5	7.5	28	10.5	0	32	0.5	0	2.5	1	3	
25258	Rijikevoort	204.81	400.21	>3	200	26.5	15.5	28	6	0	17.5	1	0	2.5	1	2	
25259	Rijikevoort	204.81	400.21	>3	200	11	5.5	40	7.5	0	27.5	0.5	0	3	0.5	3.5	
16889	Overloon	193.75	400.1	>3	100	49	8	8	7	0	0	16	0	2	2	8	
16890	Overloon	193.75	400.1	>3	100	22	14	15	14	0	0	19	0	5	2	9	

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